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A MODEL TO EVALUATE THE HYDROLOGIC RESPONSE TO LAND USE CHANGES

by

© DAVID STEVE CHANASYK

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled

A MODEL TO EVALUATE THE HYDROLOGIC RESPONSE TO LAND USE CHANGES

Submitted by DAVID STEVE CHANASYK

in partial fulfillment of the requirements for the degree of DOCTOR OF PHILOSOPHY in HYDROLOGY.

ABSTRACT

Hydrologic modelling allows watershed managers to predict the hydrologic response to watershed land use changes before they are enacted. Despite the fact that numerous models currently exist, few are capable of handling the distributed nature of such changes. In addition, existing models cannot be rigorously applied to forested, mountainous watersheds which form the head-waters of most rivers, because of the empirical treatment in these models of subsurface flow, which is the dominant flow process in such watersheds.

The objective of this treatise is to develop and to assess a hydrologic model that is simple in structure, that emphasizes the dominance of subsurface flow and that is capable of predicting the effects of land use manipulations, specifically logging, on streamflow from forested, mountainous watersheds. Development of such a model requires a novel approach since the demanding set of objectives require that the model be able to handle the water retention and water transmission characteristics of the soil profile, and also be distributed in order to simulate a wide variety of possible land use manipulations which are often imposed on only a small portion of the watershed. A model called the SLUICES (an acronym for Soils and Land Use affecting Interflow and Creating Effects on Streamflow) model is developed using this approach. The model is simple in structure, having only seven easily obtainable or optimized parameters and using a square element grid system to incorporate the distributed

nature of watershed characteristics. Choice of element size is flexible.

Simulation results for a theoretical hillslope, for various slope shapes and gradients, compared favorably with those obtained by Freeze with his more complex, more data demanding model. Both models had a soil conductivity coefficient as the most important parameter and both models showed interflow to be most significant on convex slopes.

The Jamieson Creek watershed near Vancouver, British Columbia was chosen for field calibration and validation of the SLUICES model. Research studies on this watershed have shown that interflow is the major contributor to streamflow. The model provided satisfactory simulations of measured discharges for seven test storms. Results for simulated watershed manipulations (logging) showed that the per cent increase in streamflow was directly proportional to area clearcut, as shown by field studies elsewhere. Simulations with various evapotranspiration levels explain the wide range of streamflow increases reported throughout the literature. The model showed that further success in modelling such land use changes depends primarily on the accurate assessment of evapotranspiration, particularly in the post-harvest period.

Simulations for the Jamieson Creek watershed clearly show the very dynamic, expanding and contracting nature of watershed areas exhibiting saturated soils during the course of a storm, as suggested by the contributing area concept. The simulations also show the upstream progression of soil profile saturation during a

storm. Stream lengths and drainage densities are evaluated for the Jamieson Creek watershed as variables depending on cumulative antecedent precipitation and cumulative volume of discharge.

Because of its structure and its distributed nature, the SLUICES model is flexible and can have a wide range of hydrologic applications. All of its capabilities have not been fully utilized because of data limitations, but the potential of the SLUICES model is clearly demonstrated. Other possible applications of the model are discussed and recommendations for model improvement should allow modellers to extend the model's use further.

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CHAPTER 1

LAND USE AND STREAMFLOW AUGMENTATION

With the rather paradoxical situation of having to face floods in some parts of the world and water shortages in others, man has finally begun to realize the tremendous potential influence of land use and land use changes on the behaviour of rivers. That land use has an effect had been recognized thousands of years ago, as indicated by the following statement by Plato, made about 400 B.C.[1]:

> 'There are mountains in Attica which can now keep nothing more than bees, but which were clothed not so very long ago with fine trees, producing timber suitable for roofing the largest buildings; the roofs hewn from this timber are still in existence. There were also many lofty cultivated trees, while the country produced bountiful pastures for cattle. The annual supply of rainfall was then not lost, as it is at present, through being allowed to flow over a denuded surface to the sea. It was received by the country in all its abundance, stored in the impervious potter's earth, and so was able to discharge the drainage of the hills into the hollows in the form of springs or rivers with an abundant volume and a wide distribution. The shrines that survive to the present day on the sites of extinct water supplies are evidence for the correctness of my hypothesis.'

However, it has only been in the very recent past that man has made a deliberate attempt to favorably manipulate the environment through land use practices. Modern technology is placing new and increasing demands on the environment, particularly on land and water resources. As a result,

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[1]H. C. Pereira. 1973. Land use and water resources. Cambridge Univ. Press. pp. 25 - 26.

watersheds are being subjected to many types of modifications, both minor and major, for a variety of reasons. Some modifications are designed to change the water resource capabilities of the basin, for example, to increase the watershed's water yield, to reduce flood peaks and change their timing, or to increase the recharge to aquifers. Other modifications may be carried out for purposes not directly related to water, but with probable hydrologic side effects. These may include the optimization of agricultural or forest production, attempts to increase recreational opportunities or to preserve wildlife. Larsen (1971) has delineated eighteen possible watershed modifications due to man's interference in the environment and denotes which of six possible hydrologic effects would be caused by each of the modifications.

Much scientific work has been done in an attempt to quantify the effects of deliberate manipulations of land use in many parts of the world. Work of this kind has been reported in journals such as those of the American Society of Civil Engineers, the American Society of Agricultural Engineers, the Journal of Forestry and the Journal of Soil and Water Conservation (which bears the motto 'to advance the art and science of good land use').

There is no question of the benefits of conservation, as alluded to by Plato and energetically promoted by the Soil Conservation Society of America. However, man is now beginning to face a dilemma in this regard: the dilemma of conservation versus collection. The majority of mankind lives in areas where

potential evaporation exceeds rainfall. Increased success of conservation practices in combatting soil erosion (generally by reducing the rate of runoff from fields) has resulted in reduced streamflow, which conflicts with the ever increasing demands for water downstream.

In areas blessed with more bountiful supplies of water, much can be done in the manipulation of vegetation for changing streamflows. In view of the fact that almost all of the headwaters of the major rivers are located in forested areas, it seems logical that for the purpose of streamflow augmentation, attention should be concentrated in these areas. Forested areas are unique in that they are essentially wild lands which do not receive the frequent, more intensive treatments given to croplands and grazing lands. They produce crops that are harvested at long intervals and usually it is only during harvesting that the land is subjected to deliberate disturbances. This disturbance can have profound effects upon streamflow. The net effect of forest timber removal is generally an increase in water yield.

Research in forest management has dealt for many years with the effects of fire upon both tree growth and soil erosion. The effects of logging practices and land clearing have also been studied. More recently it has been recognized that forested lands present unique opportunities for the management of water, as well as timber, as a crop. The following example will serve to demonstrate the potential that timber removal might have in this regard. In November, 1933 a forest fire denuded about 18

square kilometres of chapparal-covered mountain land in southern California. On January 1, 1934 a cyclonic storm passed over this area. Streamflow emanating from the burned out area during the storm was estimated at between 5.5 and 11 $m^3/sec/km^2$ and caused extensive damage. Meanwhile two similar but unburned watersheds nearby, which experienced the same storm, had flows of only 0.56 and 0.64 $m^3/sec/km^2$. Examinations of the watershed after the storm left little doubt that the denudation of the landscape had caused the flood (Kraebel, 1934).

Apparently the effect of vegetation upon streamflow can be profound and the manipulation of vegetation will result in the alteration of water resources. Though the evidence is quite conclusive, merely knowing the direction of the change in streamflow is not enough for watershed managers and water resource planners. Other questions follow: How will the magnitude of the peak flow be affected? Will the timing of the peak flow change, and if so, how? Will the volume of flow change? How much of the vegetation will have to be manipulated to maximize the benefits? Does the location of the manipulation matter? Will the treatments have to be continued on an annual basis? The list is long. Naturally managers and planners would feel much more confident in their decisions if these questions could be answered. The requirement for an _a priori_ decision rules out a 'wait and see' attitude, while environmental concerns rule out an 'act and see' attitude. Ever increasing public pressure demands that the environmental effects of any modifications be predicted, or at least assessed, before such modifications are undertaken in order

that undesirable effects can be minimized. These assessments have become known as environmental impact assessments. Obviously it is of compelling importance that there be a technique for evaluating the most likely effects of various land use and water management schemes, if the hydrologic effects of the modifications to the watershed are to be predicted. Simple correlations which deal with single-use management of relatively simple systems, or intuition, can no longer cope with the simultaneous changes of a large number of interacting components in the environment.

Fortunately an approach has been developed which has the potential for providing the answers to the questions posed by managers and planners and allows the evaluation of alternative plans. This approach is called hydrologic modelling. Hydrologic modelling, in the context used here, refers to the application of mathematical modelling techniques to hydrologic problems, with the aid of a computer. As pointed out by Cooper (1973) hydrologic modelling is a 'vehicle for quantifying man's impact on the environment.' The usefulness of hydrologic modelling is attested to by the great proliferation of models available for use today. Fleming (1975), in his book on simulation techniques, describes nineteen such models, with perhaps the best known being the Stanford Watershed Model and the U.S. Army Corps of Engineers Streamflow Synthesis and Reservoir Regulation (SSARR) Model.

Emphasis must be placed on the fact that hydrologic modelling has the 'potential' for analyzing the hydrologic effects

of deliberate land use manipulations for the purpose of streamflow augmentation. If the effects of any such modifications to watersheds are to be predicted, the hydrologic model utilized must be based on an understanding of the fundamental physical processes influencing the disposition of water in the watershed and be capable of expressing quantitatively these processes as they occur in the watershed. In spite of the fact that many hydrologic models currently exist, most are inappropriate for predicting the effects of proposed land use manipulations because they either do not incorporate the fundamental, physical processes occurring in the watershed, or because the nature of the model is such that it is insensitive to land use changes.

CHAPTER 2

FLOW PROCESSES IN THE HYDROLOGIC CYCLE

2.1 Examination of Possible Flow Paths

The land phase of the hydrologic cycle is often thought of as an open-ended system with precipitation forming the input and streamflow the output. Between the point of entry and the point of exit from a catchment, however, water can follow a variety of routes. The hydrograph of runoff from a basin is conveniently divided into four components which recognize these different flow paths: (a) surface (overland) flow, (b) subsurface flow (interflow), (c) base (groundwater) flow, and (d) channel interception. Figure 1 depicts the first three of these components.

Infiltration separates rainfall into surface or subsurface flow. Water failing to infiltrate into the soil forms overland flow. Water that infiltrates into the soil becomes subsurface flow. If it flows laterally through the upper soil horizons it is called interflow, but if it percolates deeper into the soil it becomes part of the groundwater flow system and may appear later as baseflow. The dominant path by which a portion of water reaches a stream depends upon such factors as climate, geology, topography, soil characteristics, vegetation and land use. In various parts of the world, therefore, one might expect that the dominant flow paths would differ.

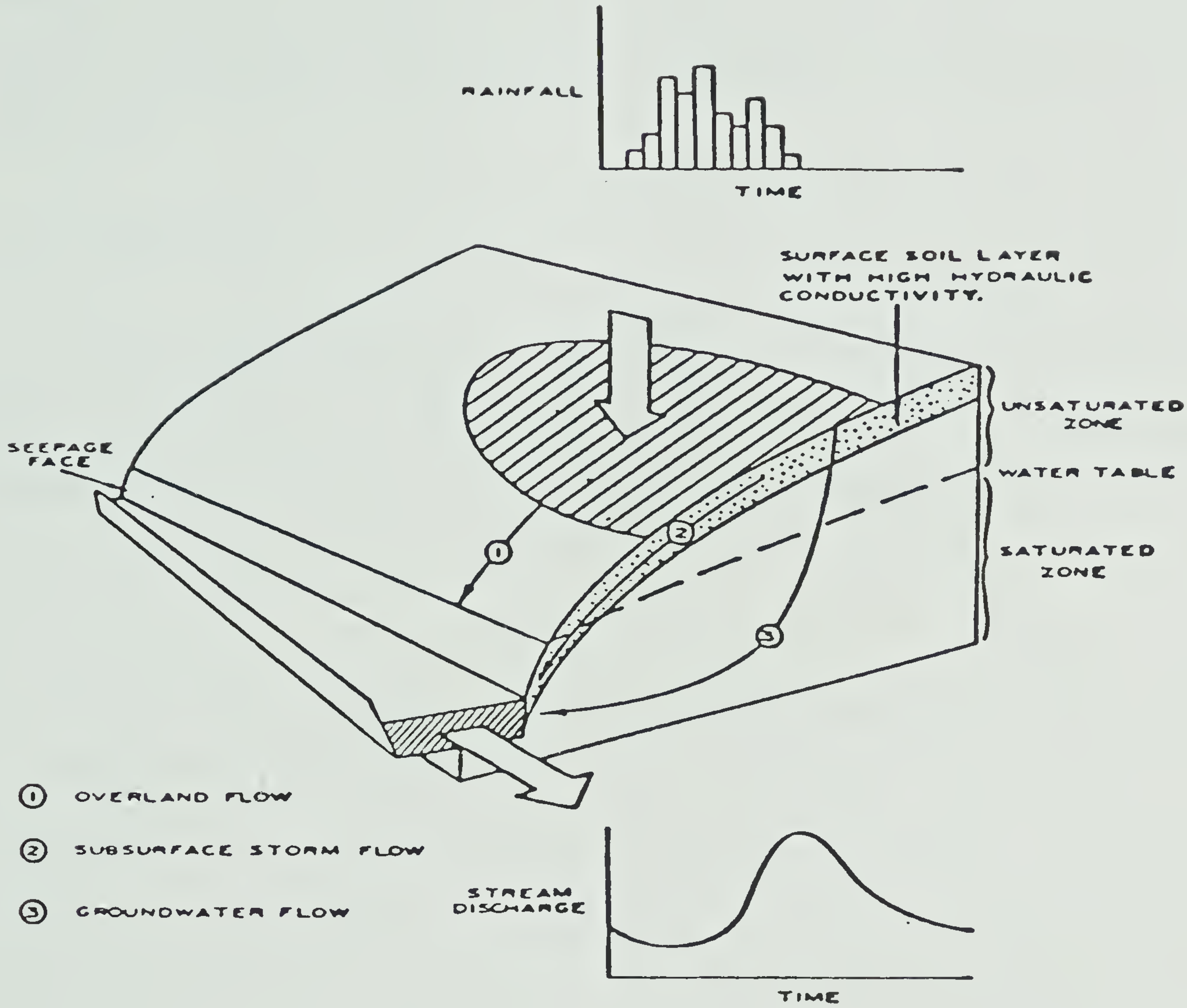

Figure 1 Hydrologic flow routes (Freeze, 1974)

2.1.1 Channel Interception

Channel interception is that portion of rainfall falling directly on the stream. If the total stream area is small, channel interception would form a relatively small portion of the total flow.

2.1.2 Baseflow

True groundwater flow (baseflow) is seldom the cause of major runoff during storms. Its primary role is in sustaining streams during the low flow periods between precipitation.

2.1.3 Overland Flow

Considerable attention in hydrologic research has been paid to overland flow. The work of R. E. Horton (1933) with its emphasis on the role of surface infiltration has had a profound influence on hydrology. Its implication is that there is a sharp demarcation between the rainfall that infiltrates the surface and the rainfall in excess of infiltration capacity, which as overland flow is responsible for all immediate runoff. The Horton model assumes that, for a prolonged storm of constant intensity, a continuous decrease of infiltration capacity occurs until a constant low value is reached. When the infiltration capacity falls below the rainfall intensity, overland flow begins. This model has received widespread recognition throughout hydrology and this concept of overland flow is widely accepted and is indeed valid for many watersheds.

2.1.4 Interflow

Where soils are pervious, where there is abundant vegetation and especially where there is a humus or litter cover, as in forested areas, the infiltration capacity is high and little direct surface runoff occurs.

The suggestion that subsurface flow is the primary source of stormflow in forested areas was made by Lowdermilk (1934), Hursh (1936), and Barnes (1939). The interflow process has been examined in much greater detail more recently in the field of forest hydrology under the leadership of Hewlett. In spite of the fact that the importance of interflow had been shown at about the same time as Horton was postulating his theory, the process has been largely ignored. This fact can be largely attributed to the uncertainty about the interflow process. Much of this uncertainty is due to the lack of a precise, widely accepted definition for interflow even to this day. Hursh (1936) first called it subsurface stormflow or storm-seepage and defined it as the stormflow which infiltrates into the surface soil but moves away from the area through the upper soil horizons at a rate much in excess of normal ground water seepage. Barnes (1939) accepted the term and extended the definition to include water which had penetrated only the upper soil layers during a rainstorm or a thaw and has filtered more or less horizontally through the soil to discharge into the stream system by seepage. It appears that it was Barnes who, recognizing the confusion surrounding this particular flow process, proposed the term

interflow, which is a fairly widely accepted term appearing in most hydrology textbooks. Another term which is sometimes used synonymously with interflow is throughflow. Freeze (1972b) defined subsurface flow (interflow, throughflow) as that part of lateral flow that infiltrates the soil surface and moves laterally through the upper soil horizons toward the stream channel as unsaturated flow or as shallow, perched saturated flow above the main groundwater table.

There are two common features to the three definitions presented above: (1) Interflow results from water that has infiltrated into the soil, and (2) Interflow is mainly a lateral flow process in the upper soil horizons. These similarities might suggest a universal agreement among researchers, but examination of the literature suggests that controversy arises when the importance of interflow is assessed in relation to other flow processes. Misinterpretation is largely due to the lack of an appreciation of the nature of the interflow process and to the assumption that interflow is water that moves towards the stream and discharges there completely independently of either surface runoff or groundwater flow: that is, there never is any interaction between the processes. This possible intermingling of flows has important repercussions on the assessment of the role of interflow. If intermingling does occur, discharge of water that began as interflow could be of the overland flow type at the stream. Though the contribution of interflow would be masked, it could nonetheless be important. A definition that takes account of this was proposed by Gray (1970): Interflow is water

infiltrating the soil surface and moving laterally through the upper soil horizons until it is intercepted in its course by a stream channel or returns to the surface (at some point downslope from its point of infiltration) to flow to the stream as surface runoff. This definition will be adopted.

2.1.5 Discharge Hydrographs

The particular flow processes experienced on a watershed are reflected in the discharge hydrograph. Differences in hydrograph shape are largely due to the dramatic differences among the velocities of flow of the processes. Overland flow occurs at the quickest rate with groundwater flow being extremely slow, the two rates differing by as much as several orders of magnitude. Interflow occurs at a rate somewhere between those of overland flow and groundwater flow. Overland flow produces a hydrograph whose form is largely determined by the duration of the storm, the variation in its intensity and by the storage characteristics of the stream channel. The hydrograph of interflow displays a flattened peak and a much more gradual recession than that of overland flow, but can show rapid response. Its shape depends mostly on the near surface soil characteristics of the watershed. The hydrograph of baseflow always shows a very conspicuous time lag between the storm and the rise in flow, with the recession being very gradual.

Figure 2 (Freeze, 1972b) depicts the variety of hydrograph shapes that might be commonly experienced. Hydrograph (a) would most closely represent the response to Hortonian overland

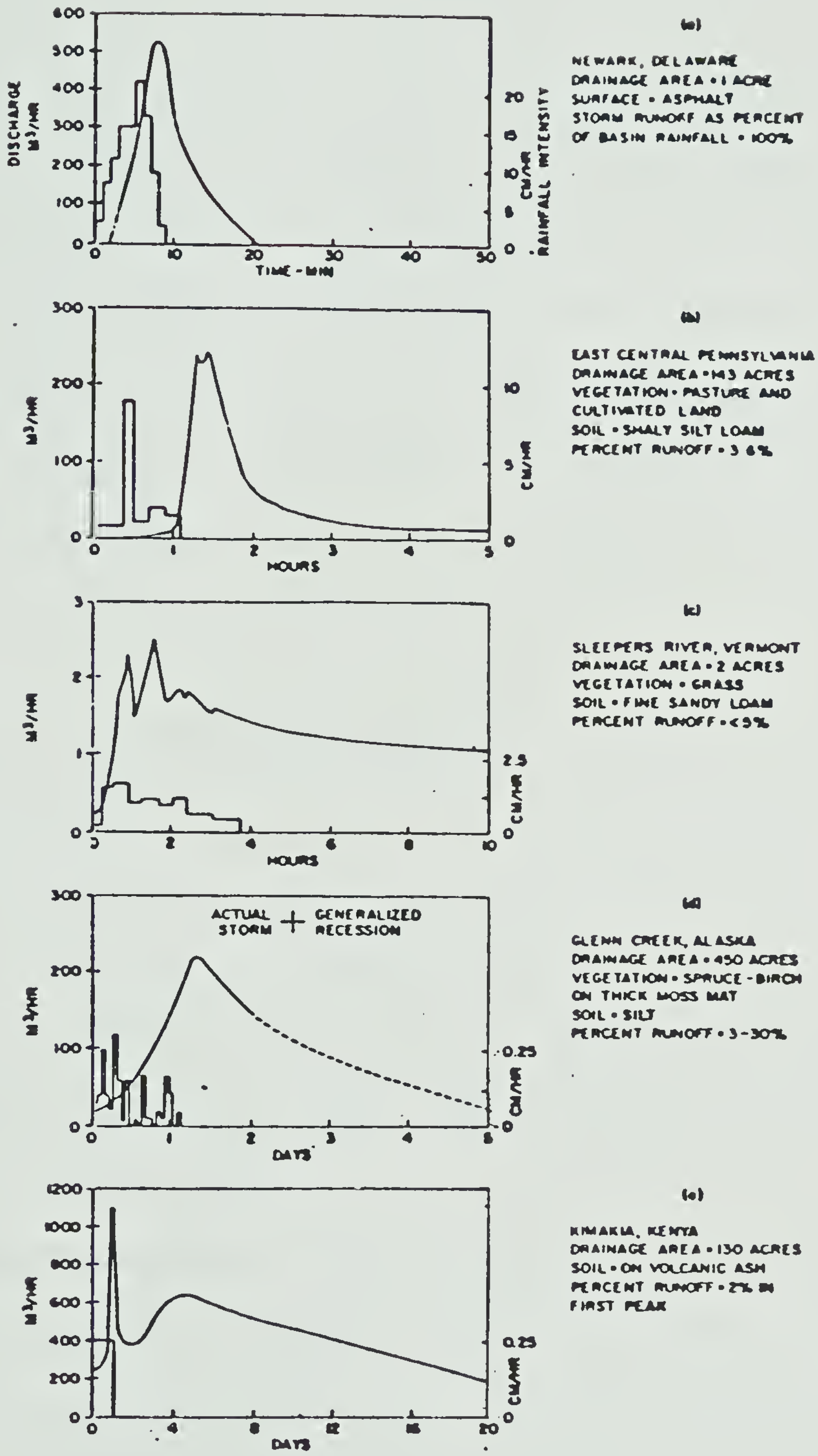

Figure 2 Recorded hydrographs for small watersheds (Freeze, 1972b)

flow as it represents the discharge from an asphalt parking lot. Hydrograph (b) is the result of subsurface flow since it was recorded for a storm in which no overland flow was observed. Hydrograph (c) depicts the great sensitivity of channel discharge to rainfall intensity. Measurements at the site did not confirm whether or not the prominent recession limb was due to later subsurface contributions. Because hydrograph (d) was taken from a watershed in Alaska where a thick peat moss layer overlays a relatively impermeable silt layer; it would represent a slow subsurface storm response. Hydrograph (e) shows a secondary peak caused by base flow. This type of response is generally quite rare.

2.2 The Role of Interflow

2.2.1 The Importance of Interflow

In the past fifteen years interflow has been shown to be a major, if not the dominant component of total stormflow from vegetated drainages (Whipkey, 1965, 1969; Kirkby and Chorley, 1967; Hewlett and Hibbert, 1967).

Whipkey (1965) suggested that interflow occurs especially where: (1) the land is sloping, (2) the surface soil is permeable, (3) a water impeding layer is near the surface, and (4) the soil is saturated. A significant presence of interflow in certain watersheds has been demonstrated by isotope techniques with the proportion of subsurface flow found to exceed 50% in field experiments (Dincer et al., 1970; Martinec, 1975; and

Holocek and Noujaim, 1975). The importance of subsurface flow was also confirmed by a field study in a well-instrumented area by Stephenson and Freeze (1974).

Quick watershed response has usually been attributed to the classical concept of streams being fed by overland flow generated on partial areas of the basin. Stephenson and Freeze (1974) point out that this mechanism is rarer in both time and space than had previously been supposed. Hewlett and Hibbert (1967) suggest that for forested land the most logical approach is to assume that all flow is subsurface until there is evidence to the contrary.

2.2.2 Field Studies of Interflow

Numerous field experiments have been undertaken to gain an understanding of interflow, particularly on sloping land. Several of these studies, for the period 1944-1977 are reviewed below. The works of Hewlett (1961), Hewlett and Hibbert (1963), and Nutter (1975) were excluded from the review because of the homogeneity of the artificial soil mass that they considered.

2.2.2.1 Hursh and Hoover (1941) utilized a field installation in the southern Appalachians for measuring storm water movement at various depths in a forest soil profile (see Appendix I for soil profile characterization). The installation involved digging a 2.4 m by 2.4 m pit on a 20% slope and then placing horizontal catch troughs across the exposed soil profile face. One trough was set at the soil surface, another at a depth of 30 cm (more

significantly at the bottom of the A horizon) and another at the base of the pit approximately 90 cm down. Storm runoff measurements from the plot were taken after the plot had been brought up to field capacity. Results from 1 cm of precipitation over a 15 minute period showed that 2 1/2% of the precipitation occurred as surface runoff, 12% moved below the surface in the A horizon, with the rest being accounted for below this level.

2.2.2.2 van't Woudt (1954) studied interflow in volcanic ash soils in New Zealand on a 17 m long virgin country site under shrub, with an average slope of 62%. The soil exhibited both an A horizon and a B horizon, the latter extending to a depth of 1 m at the foot of the slope. The investigation involved using small lysimeters, placed in the soil at depths ranging from 15 to 60 cm, along the length of the slope which were capable of trapping water flowing laterally from higher ground. Interflow was found to take place as a result of the obstruction of percolating water and took place through a relatively thin layer of wetted soil at the surface. The lower infiltration capacity of the B horizon in comparison to that of the A horizon, by a factor of 2, caused a buildup of moisture in the A horizon, resulting in lateral flow through the layer. A similar type of moisture buildup and lateral flow was caused by soil layering in cases where a fine-textured soil overlays a coarse-textured soil.

2.2.2.3 Whipkey (1965) utilized a site in east-central Ohio having a sandy loam soil and a slope of 28%. The vegetative cover of the study area was a mixed oak stand and the plot was

covered with a 5 to 10 cm depth of mixed hardwood leaf litter. The soil was very permeable down to a depth of 90 cm, showing a slight decrease in vertical hydraulic conductivity at 56 cm. The plot itself was 17 m long, 2.44 m wide and was oriented down the slope. Troughs to collect seepage water from the slope face were located at the surface and at depths of 56, 90, 120 and 150 cm (essentially at the separation points of the major soil layers). Suction head measurements were made using tensiometers and moisture content measurements were made with a neutron probe. Results for all storms showed that 64% of all flow came from the 56 - 90 cm layer (the layer above the flow impeding layer) with approximately equal amounts passing through the 0 - 56 cm layer and the 90 - 120 cm layer (16%). The deepest layer, 120 - 150 cm, showed only slight flow (3%) and surface runoff was even less, accounting for only 1% of the flow. Tensiometer readings showed that a saturated layer existed above the plane of the wetting front.

2.2.2.4 Dunne and Black (1970) chose a 0.24 ha hillside in north-eastern Vermont as their study area. At the time of the study, the hillslope was in pasture, but had been covered by a pine forest until fire destroyed it some thirty years prior to the study. The hillside had a southerly aspect with the slope varying from 30 to 100%, with a relief of 18 m. It was divided into three plots: one with a convex profile and an area of 0.05 ha, one with a concave profile and an area of 0.12 ha, and one whose profile was straight with an area of 0.07 ha. The soil

texture of the slope was generally sandy loam. Runoff was intercepted by drains in a long trench spanning the three plots and running adjacent to the stream. The trench was 1.5 to 3 m in depth, depending upon the depth of the till material. Runoff was collected at the soil surface, the base of the root zone (30 - 75 cm) and at the groundwater seepage level. Soil moisture was measured with a neutron probe. Water table fluctuations were also noted. Both artificial and natural storms were studied on only the concave and the straight slopes. Hortonian overland flow was not experienced at anytime on the study area as the infiltration capacity exceeded rainfall intensity at all times. However, overland flow was measured. This flow was caused by a combination of water emerging from the saturated soil and rainfall that had fallen directly onto this saturated area. The authors noted that overland flow of this type originated on a small concave portion of the hillside, comprising 5 to 10% of the concave and straight plot areas.

They concluded that only when water was released from the damping effect of unsaturated flow by emerging at the ground surface could it contribute to channel runoff at a significant rate. They considered subsurface flow from the hillside as being relatively unimportant, contributing only a small part of the water causing the stream hydrograph peak. They rationalize by stating that the return flow would have velocities 100 to 500 times greater than those of subsurface flow, allowing more water to travel from a larger contributing area in the time available. The authors concluded that the importance of a hillside as a

producer of storm runoff depends upon its ability to generate overland flow.

2.2.2.5 Weyman (1973) studied a hillslope in Great Britain 670 m long, convex in profile with a relief of 50 m. The slope shape was complex, ranging from 42% at the base to 27% at midsection to 3% near the divide. Four soil layers were identified, the boundaries of which were located at depths of 10, 25, 40 and 60 cm. Discharge was measured from each of four layers, for a width of 1 m, using lateral troughs. Results showed that no overland flow occurred. All runoff from the hillslope was subsurface flow, confined entirely to the two lowest soil horizons, the B and the B/C horizons. Weyman suggested that water leaves the soil through a zone of permanent soil saturation at the slope base and that during the course of a storm, this zone of saturation grows upslope and up the soil profile in the form of an expanding wedge. He noted that this saturated zone is fed by water moving out of the unsaturated zone lying above and upslope of the saturated zone. He postulated that the unsaturated percolation following substantial rainfall overloads the lateral unsaturated flow system of the B/C horizon, causing water to accumulate at the profile base until saturated conditions are produced. Under heavy infiltration, this process could initially occur at the base of the B horizon, resulting in a temporary perched saturated zone overlying an unsaturated zone. Discharge might then occur from the B horizon before it occurred from the B/C horizon.

Weyman noted that the initiation of saturated lateral flow is dependent upon some break in the vertical hydraulic conductivity profile of the soil, which in this case occurred both at the base of the B horizon and at the soil profile base. Thus the primary controls on hillside response were the depth to such changes in the vertical hydraulic conductivity of the soil and the velocity of lateral flow. Weyman, in conclusion, doubted whether subsurface flow, considering the slowness of this type of flow, could significantly contribute to actual stormflow and suggested that overland flow would still dominate the storm response of most drainage basins. He did point out, however, that subsurface flow would form a very important element of base response during the hydrograph recession.

2.2.2.6 Harr (1977) studied subsurface flow in a forest soil on a 10.23 ha watershed in Oregon. The study area was located on a stream to ridge portion of the watershed, with a slightly convex, complex slope ranging from 50% near the ridge to 110% adjacent to the stream. The total relief was approximately 105 m. Only slightly more than 1 m of poorly developed soil existed above 2 - 7 m of subsoil. Notable changes in saturated hydraulic conductivity values occurred at depths of 70 and 130 cm. Results showed that channel interception and subsurface flow averaged 38% of gross precipitation, ranging from 23 to 51%. Channel interception was never more than 4% of total flow. Interflow was therefore the major contributor to storm runoff. Piezometer data showed that saturated zones expanded upslope

and laterally as rainfall continued. There was no evidence to suggest that overland flow on wet zones near the stream had ever occurred. The dominance of interflow can be explained by the high soil hydraulic conductivities and steep, convex slopes, which would be conducive to interflow, and the fact that saturated zones frequently occurred at the soil subsoil boundary at midslope to upslope locations.

2.2.2.7 Summary

The above review of six experimental field studies of interflow on hillslopes has shown that interflow can be an important flow process. However, controversy about the importance of interflow relative to the other flow processes, most notably overland flow, is evident. Any study that attempts to investigate the effects of land use changes on streamflow in small forested watersheds must first quantitatively assess the role that interflow plays in generating streamflow in these watersheds. This assessment should be based on a clear understanding of interflow as a physical process, how it forms and how quickly it can respond to precipitation. This understanding requires knowledge of the soil's water storage and water transmissive capabilities.

CHAPTER 3

THE SOIL PROFILE AS A RESERVOIR AND A CONVEYOR OF WATER

3.1 Water Storage Characteristics of the Soil

Soil has two basic components: solid soil grains and void space. The void space of the soil can be occupied by either air or water and is characterized by porosity. The amount of water occupying the voids can be conveniently expressed as a volume percentage moisture content, which is the ratio of the volume of water to the total soil volume. Three main types of soil water can be identified: hygroscopic water, plant available water and gravity water. Hygroscopic water is the water remaining below wilting point; plant available water is that water held between field capacity and wilting point; and gravity water exists when the moisture content exceeds field capacity and can be drained by the force of gravity. Wilting point represents the point below which plants cannot extract water. Field capacity is the moisture content of the soil after gravitational drainage of soil water has significantly decreased. In principle, particularly in soil with a high clay content, the amount of hygroscopic water may be appreciable. However, it is generally disregarded because it is unavailable for plant use or for drainage. Thus, practically, the water content between wilting point and saturation is most important.

The volumes of solid material and the total void space will vary from soil to soil, depending upon soil texture and structure. The typical compositions of sand and a clay soil show striking differences:

(1) A clay soil generally has a greater total porosity than a sandy soil.

(2) A clay soil has a much greater amount of hygroscopic water.

(3) A sandy soil has a larger portion of gravity water.

(4) The ratio of plant available water to gravity water is usually greater for a clay soil than for a sandy soil.

The magnitudes of water content at the saturation level, field capacity and wilting point are important, but the difference between the values is of greater consequence in determining the mobility of the soil water. This is verified by the fact that, in spite of having lower porosities (lower overall water storage ability), sandy soils can supply greater volumes of drainage water because their gravity water component is greater than that for a clay soil. Thus, in spite of the fact that porosity is indicative of the water storage potential of the soil, by itself it gives no indication of the water available for drainage or for plant use. Therefore, it is important to subdivide the total storage into the three types of soil water. This can be done by specifying the values of field capacity and wilting point.

If the premise that the major zone of hydrologic activity lies above some depth which controls profile drainage is accepted, the storage volumes associated with the various water types can be calculated. This control depth is usually

characterized by a sharp increase in bulk density and a decrease in porosity. The storage potential of the soil profile, after the start of precipitation, is reduced by infiltrated volumes of water. During periods of no precipitation, there is a recovery of available storage through the drainage of gravity water and the evapotranspiration of plant available water. For practical purposes, drainage and evapotranspiration are assumed to become negligible at field capacity and wilting point respectively. In watershed management, these water storage volumes are of importance because the evapotranspiration by vegetation plays a very important role in determining antecedent conditions and thus the amount of runoff from the watershed. The water holding capacities of soils can be approximated from information on soil texture. Commonly the range of values normally experienced for any given texture is rather narrow.

England (1970) presented the following values for the respective soil moisture storages for various soil textures:

Soil Texture	Gravity Water (cm/m)	Available Water (cm/m)
coarse sandy loam	15.8	8.7
sand	19.0	13.3
loamy sand	26.9	10.1
sandy loam	18.6	12.3
fine sandy loam	23.5	13.1
loam	14.4	15.6
silt loam	11.4	19.9
sandy clay loam	13.4	11.9
clay loam	13.0	12.7
silt loam	8.4	14.9
sandy clay	11.6	7.8
silty clay	9.1	12.3
clay	7.3	11.5

Use of this table would allow one to estimate soil moisture storages based only on information about soil texture.

3.2 Water Conductive Properties of the Soil

Darcy's law states that the velocity of flow of a fluid through a porous medium, expressed as discharge divided by gross area, is proportional to the hydraulic gradient causing the flow, with the proportionality factor being the hydraulic conductivity of the medium. Because for given fluid properties, hydraulic conductivity is a characteristic physical property of the medium, it should be related to measurable properties of the soil pore geometry. The Hagen-Poiseuille equation, describing laminar flow through a capillary tube, demonstrates the importance of capillary (pore) size to flow. Thus, the hydraulic conductivity of a porous medium depends not only on the total pore space of the soil but also on the size and shape of the pores, as well as the viscosity and density of the liquid flowing through the soil.

The value of hydraulic conductivity varies with moisture content. It is a maximum when the soil is saturated and declines to near zero in dry soil. When the soil is saturated, essentially all of the pores are filled and conducting fluid. When the soil is unsaturated, some of the pores are air-filled, and the conductive portion of the soil's cross-sectional area is decreased. Furthermore, as moisture content decreases, the first pores to empty are the largest ones. This forces the water to flow through the smaller pores, which are much less conductive. For

these reasons, the transition from a saturated to an unsaturated medium causes a steep decline in hydraulic conductivity, usually several orders of magnitude, as moisture content decreases. At saturation the most conductive soils are those in which large and continuous pores constitute most of the overall pore volume. Thus, a sandy soil exhibits a higher hydraulic conductivity than a clayey soil at saturation. However, the very opposite may be true when the soils are unsaturated. In a soil with small pores, such as a clay, many of the pores will remain conductive even at low moisture contents so that the hydraulic conductivity does not decrease as steeply as it would in a sandy soil as moisture content decreases. Hydraulic conductivity may actually be greater for the clayey soil than for a sandy soil, under these unsaturated conditions.

Because of this complex behaviour, there is no reliable way of predicting hydraulic conductivity from intrinsic soil properties: it must be measured experimentally. However, it may not be always advisable to determine the hydraulic conductivity of the soil in question in the laboratory using artificially repacked samples or undisturbed cores and then to estimate moisture movement in the field using these values of hydraulic conductivity. Also the values of hydraulic conductivity determined for small samples in the laboratory may not be representative of larger masses of soil in the field, where fissures and cracks may have a significant effect on water transmission. Field techniques, though probably providing the best value of hydraulic conductivity, remain tedious and time

consuming. The problem of representativeness of field samples must also be reckoned with. Hence, interpretation of experimental measurements can pose many problems. However, because of the importance of such interpretations, reported values of hydraulic conductivity from various sources should be compared.

Whipkey (1965) reported the following results for a site at the Central States Forest Experiment Station at Columbus, Ohio:

Soil Depth (cm)	Textural Class	Density (g/cm^3)	Hydraulic Conductivity (cm/h)
0 - 56	Sandy Loam	1.33	-
56 - 90	Sandy Loam	1.41	28.6
90 - 120	Loam	1.78	1.7
120 - 150	Clay Loam	1.80	0.2

An impeding layer appears to begin at a depth of 90 cm. At this level, the bulk density increases suddenly and the hydraulic conductivity decreases. The general trend of decreasing hydraulic conductivity with increasing depth in the soil profile is also evident.

Harr (1977) reported the following results for a field site with a clay loam soil profile:

Depth (cm)	Bulk density (g/cm^3)	Porosity (%)	Saturated hydraulic conductivity (cm/h)
10	0.807	60.8	352
30	0.897	63.8	412
70	1.015	60.3	163
110	0.981	63.1	175
130	1.080	55.4	16
150	1.053	57.6	22

The soil profile was 1 m in depth. Harr suggests that differences in soil aggregation most likely account for differences in hydraulic conductivity.

Dunne (1978) presented a compilation of measured values of saturated hydraulic conductivity of soils in which interflow has been measured. This compilation is presented in Table 1.

3.3 Effects of the Nonhomogeneity of the Soil Profile

Undisturbed forest soil is generally covered by an organic litter that protects the soil surface and keeps it permeable to water infiltration. In addition, the A and B horizons of forest soils are interlaced with roots, animal burrows, and structural channels that provide a highly permeable medium for the rapid movement of water in all directions.

There are few soils in the field which have uniform texture and structure throughout the profile. The physical properties of the respective horizons within a given profile can vary tremendously. The contrast between horizons in forest soils can be dramatic and the horizons, particularly the B horizon, may be thicker on level land or at the foot of the slope than on sloping land or at the crest of a slope. In nature, however, sharp contrasts in soil properties are unusual with gradual changes being a more normal feature. The normal sequence of soil horizons down from the surface, from litter to the partly organic A horizon to the B horizon is characterized by a decrease in porosity and often by an increase in clay content. Bulk density commonly increases with depth as well. This is a

Table 1 Saturated Hydrologic Conductivity of Soils in Which Subsurface Stormflow Has Been Measured *

Soil type	Saturated hydraulic conductivity (cm/hr)	Source
Upper 7·5 cm of a sandy loam (A_0 horizon)	118 (highest of a series of measurements)	Laboratory measurement, Dunne (1969a)
Sandy loam topsoil	34·2–37·2	Field measurements, Dunne (1969a)
Sandy loam	30·5	Field measurements, Hewlett and Hibbert (1963)
Sandy loam (56–90-cm depth)	28·6	Field measurement, Whipkey (1965)
Sandy loam (7·5–60-cm depth)	Mean 24·3 Range 17·2–46·0	Laboratory measurements, Dunne (1969a)
Silt loams and loams	Median 8·4–10·4 Range 0·15–16·5	Field measurements, Rawitz *et al.* (1970)
Varved sandy silt subsoil	Mean 8·9 Range 1·3–18·5	Laboratory measurements, Dunne (1969a)
Varved sandy silt subsoil	Mean 4·8	Field measurements, Dunne (1969a)
Clay loam topsoil	2·5–7·5	Field measurements, Betson *et al.* (1968)
Loam subsoil (90–120-cm depth)	1·7	Field measurement, Whipkey (1965)
Clay loam subsoil	0·75	Field measurements, Betson, *et al.* (1968)
Clay loam subsoil (120–150-cm depth)	0·2	Field measurement, Whipkey (1965)

* Dunne (1978)

reflection of the decreasing porosity of the soil and will cause a decrease in hydraulic conductivity with depth. Because the soil water must move through the pore space of the soil, any change in texture or structure will result in a change in the water transport and storage properties of the soil.

The moisture retained in a soil is determined by the characteristics of the whole profile, which may include several distinct horizons. The texture of underlying layers can also play an important role in this regard. In general, if the wetting front encounters a layer in which most of the pores are either larger or smaller than those through which it is moving, an effect on the rate of flow will be observed.

When a wetting front contacts a material with finer pores than that in which it has been moving, the fine pores begin to fill rapidly because of their greater attraction for water. As the wetting front advances into the material, water must be transmitted through the fine pores which have filled with water. If the material has extremely fine pores, such as those found in clayey soils, resistance to flow because of the fineness of the pores may be so great that flow is markedly reduced. Colloidal swelling which occurs in many fine materials may further reduce the size of the transmitting pores, and decrease flow. This reduction in flow rate can result in the formation of a saturated layer in the more pervious layer. Figure 3a shows the theoretical pressure distribution with depth for a two layer soil profile with the hydraulic conductivity of the first layer exceeding that of the second. Note the buildup of positive pressures at the

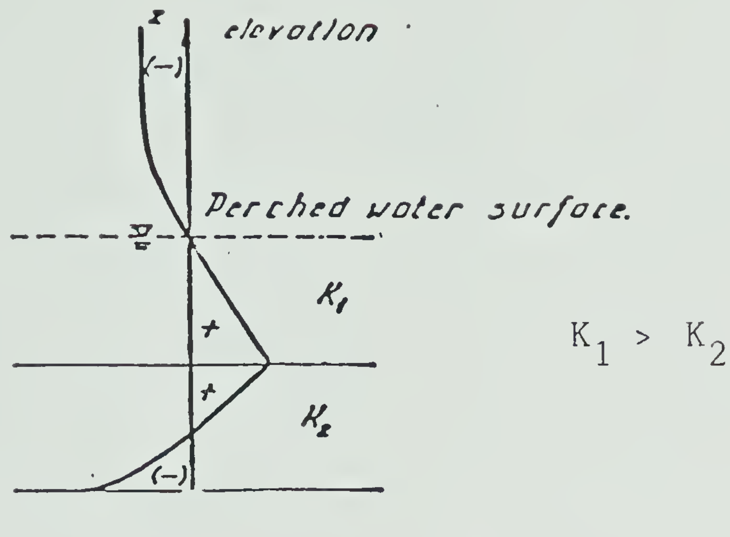

Figure 3a Theoretical pressure diagram for a two-layered system (Zaslavsky, 1964)

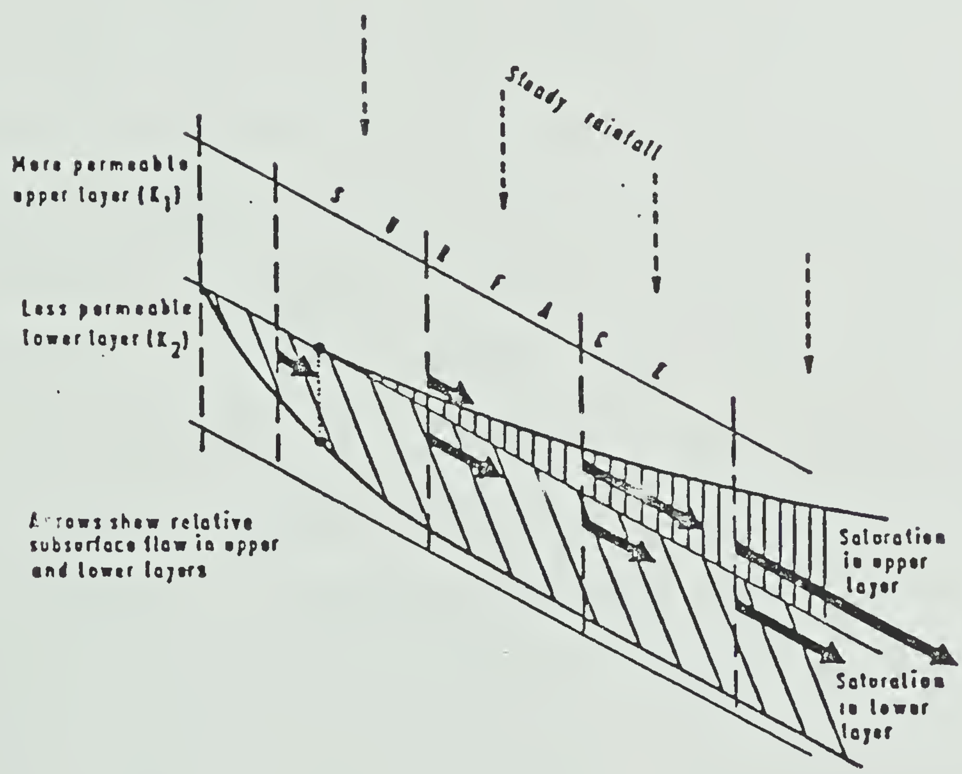

Figure 3b Formation of a saturated layer at an impeding layer (Whipkey and Kirkby, 1978)

interface of the first and second layers. The formation of a perched water surface in the uppermost layer indicates that the second layer is acting as a flow impeding layer. Figure 3b shows a hillslope characterized by a permeable upper layer underlain by a less permeable layer. A wetting front (saturated layer) is shown moving into the second layer. This front would advance vertically downwards as well as laterally. The saturated layer in the upper more permeable layer will create interflow due to the slope of the layers. This flow will be predominantly in a lateral, downslope direction.

Data reported by Day and Luthin (1953) support this theoretical derivation for the formation of a saturated layer at a profile discontinuity. A layered column consisting of a highly permeable stratum (Yolo very fine sandy loam) overlying a less permeable stratum (Yolo loam) was examined. Water was added to the surface and maintained at a constant head while the water was allowed to escape out the bottom. The resulting relationship between pressure and depth is shown in Figure 4, demonstrating the formation of a positive pressure zone (saturated layer) at the interface. The pressure increased steadily with depth from the surface to a maximum value near the contact plane and decreased thereafter to atmospheric pressure at the outflow point.

When a wetting front moving in a soil characterized by relatively fine pores contacts material with larger pores, the volume of pores in this latter layer capable of holding water under the same pressure conditions decreases. Before the wetting front can advance these larger pores must fill with water. The

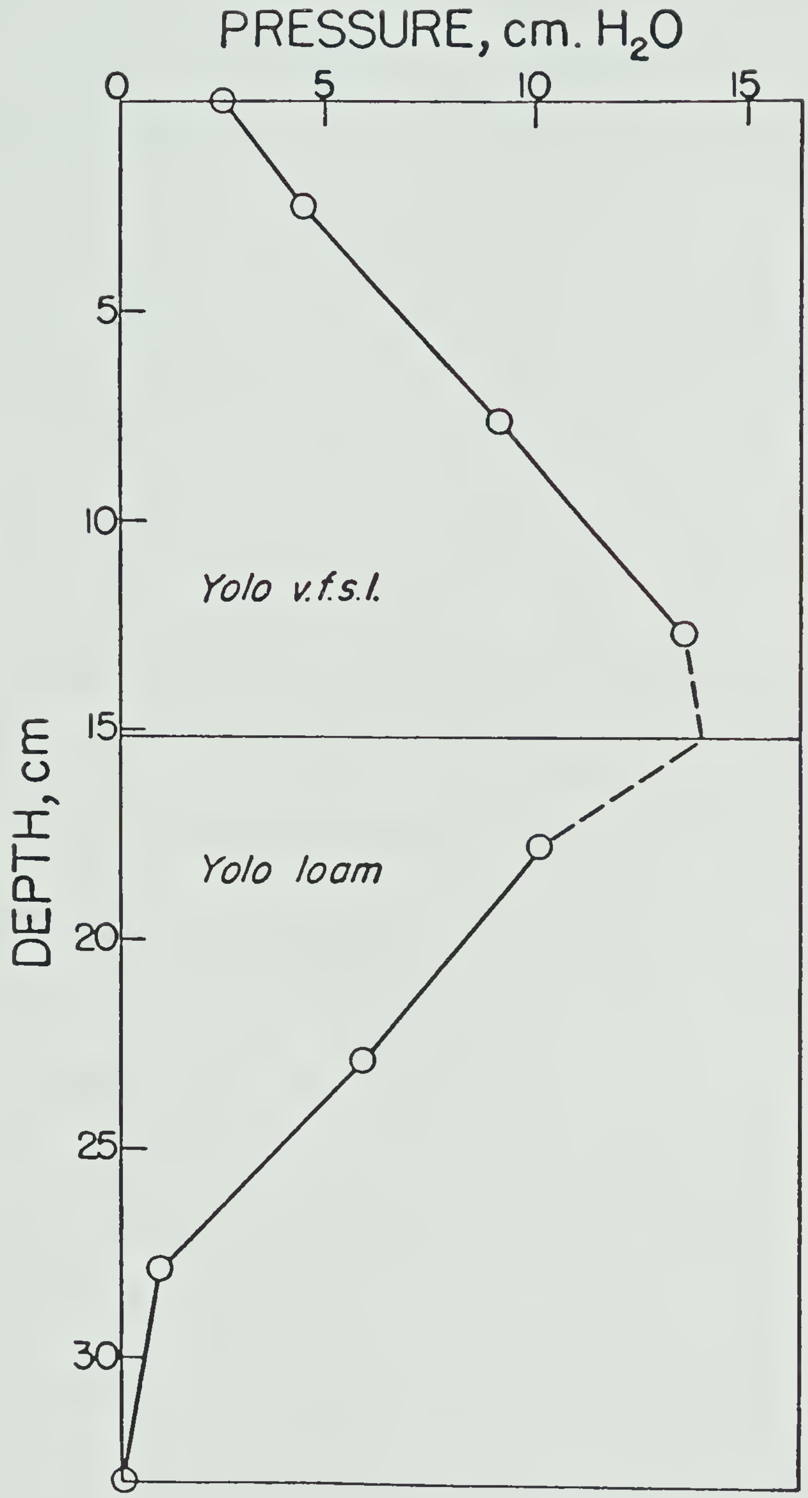

Figure 4 Experimental verification of pressure buildup at a soil discontinuity (Day and Luthin, 1953).

time delay for equilibrium to be reached causes a restriction in the advance of the wetting front and lowers the flow rate. Thus a coarse-textured layer underlying a fine-textured layer can increase the storage capacity of the finer layer (Alway and McDole, 1917). This water could then form interflow.

Because a soil profile is made up of several horizons, opportunities for flow impediment and thus soil saturation exist at each horizon - horizon interface. Whipkey and Kirkby (1978) suggest that in a multi-layered soil the most critical layer with respect to water transmission is the one with the lowest saturated hydraulic conductivity in the soil profile. The second most critical layer is the layer which has the next lowest hydraulic conductivity and lies above the most critical layer, and so on. In this way, a sequence of critical soil layers may be established. Each layer could potentially act as an impeding layer if the percolation rate of the layer above it exceeds its saturated hydraulic conductivity. There are many variations of profiles which have less permeable subsoils. Slowly permeable subsoil layers may have formed in place or may have resulted from an earlier deposition of soil material, such as lacustrine deposits that have later been covered with windblown soil. An extreme case is soil underlain by unfractured bedrock.

The data presented by Harr (1977), discussed earlier, provides an excellent demonstration of the concept of the impeding stratum. Note particularly the changes in porosity and saturated hydraulic conductivity at depths of 30 and 110 cm. The rather distinct changes suggest important horizonation at two

levels. Since the change in saturated hydraulic conductivity between depths of 110 and 130 cm is so dramatic (more than an order of magnitude), one could conclude that the impeding layer lies at a depth of 110 cm. The change in saturated hydraulic conductivity at this depth would prove to be an impediment to moisture flow. Beke (1969), in a study of the soils of Marmot Creek Basin in Alberta, found that most of the soils had an impeding layer within 30 cm of the mineral surface, with this layer generally being characterized by an increase in clay content.

Good approximations for the depth to the flow impeding horizon can generally be obtained from a soil profile description. Abrupt changes in texture, consistency or structure are generally indicative of changes in water transmissibility. Refined estimates to the depth of the impeding layer can be obtained from laboratory determinations of bulk density.

The saturated layer of water that forms above a flow impeding layer may be receiving not only percolating water from the soil surface, but if the soil is on a slope, from upslope as well. If the influx of water to a zone having a saturated layer continues at a sufficient rate for long enough the saturated layer may build up to the soil surface. If the whole profile becomes saturated, overland flow will be produced. The ability of the profile to become saturated will depend on the depth to the impeding layer as well as the water holding capabilities of the soil above this horizon.

CHAPTER 4

THE MATHEMATICAL CHARACTERIZATION OF INTERFLOW

Darcy's law gives the relationship between velocity of flow in a porous medium and the hydraulic gradient. This law, though originally conceived for saturated flow, was extended by Richards (1931) to unsaturated flow, recognizing that the hydraulic conductivity becomes a function of moisture content.

Using Darcy's law and the continuity equation, the following generalized equation of flow in three dimensions can be derived (Freeze, 1969):

$$\frac{\partial}{\partial x}\left[\rho K(x, y, z, h_t)\frac{\partial h}{\partial x}\right] + \frac{\partial}{\partial y}\left[\rho K(x, y, z, h_t)\frac{\partial h}{\partial y}\right] + \frac{\partial}{\partial z}\left[\rho K(x, y, z, h_t)\frac{\partial h}{\partial z}\right] = \rho\frac{\partial \theta}{\partial t} + \theta\frac{\partial \rho}{\partial t}$$

with x, y = horizontal coordinates,
z = vertical coordinate ,
ρ = density of the fluid,
K = hydraulic conductivity,
h_t = suction head,
h = hydraulic head,
θ = volumetric moisture content,
t = time.

As Freeze suggests, this equation can be used to develop the one-, two- or three-dimensional forms of the steady or unsteady flow equations for saturated or unsaturated flow of a compressible or incompressible fluid through a homogeneous or nonhomogeneous medium. For a particular characterization of fluid flow in porous media, certain assumptions can be made to

simplify this general, but complex, equation. For example, if the fluid is assumed to be incompressible and the medium to be homogeneous and isotropic, the well-known Laplace equation can be derived from the above equation. Unfortunately, the presence of soil horizons suggests that the Laplace equation would not be applicable to most field soils, which by their inherent nature are both nonhomogeneous and anisotropic.

4.1 Study Disciplines

Quantitative studies of the movement of water in saturated or unsaturated soils have been made in various disciplines, such as soil science, civil and petroleum engineering and hydrogeology, but have developed largely independently.

4.1.1 Soil Science

Soil scientists, being concerned with the availability of water to plants, have worked primarily in the unsaturated zone. Usually the water table is assumed to be at a deep, insignificant depth, thus avoiding the possible contributions of a capillary fringe emanating from it. A few broad categories of study include water movement below the root zone, water uptake, evapotranspiration and infiltration. Due to the relative flatness of agricultural land, the flow processes under consideration are essentially one dimensional, vertically. This simplifies the flow equation considerably, though problems of nonhomogeneity and hysteresis (the change of physical properties depending upon the wetting and drying history of the soil) still remain. For

unsaturated, unsteady incompressible flow the general equation reduces to:

$$\frac{\partial}{\partial z}\left[K(h_t)\frac{\partial h}{\partial z}\right] = \frac{\partial \theta}{\partial t}$$

Philip (1957) developed a numerical procedure which enables solution of this equation for vertical infiltration into a homogeneous semi-infinite medium with a constant initial moisture content. Hanks and Bowers (1962) presented a numerical solution, using finite difference techniques, to the cases of vertical and horizontal infiltration. The solution technique requires, in addition to a knowledge of the initial and boundary conditions of the specific problem, a known relationship between moisture content and suction head as well as one between moisture content and hydraulic conductivity. The technique has distinct advantages in that it does not require that the soil be homogeneous or semi-infinite, that gravity be neglected or that the initial moisture content be uniform. Complications due to hysteresis would still limit application of the technique to problems where the relationship between moisture content and suction head are known. A noteworthy conclusion of the authors' layered experiments was that vertical infiltration was governed by the least permeable layer, once the wetting front reached this layer.

The more general multi-dimensional flow problems have not as yet undergone as comparable an extensive treatment as have the unidimensional ones, even though the former are of considerable importance in hydrology. The introduction of the

computer has allowed the use of numerical methods, including finite difference and finite element techniques in the solution of problems that more closely simulate field conditions, but which, because of soil heterogeneity and anisotropy, will be very complex.

4.1.2 Geotechnical Engineering

Investigations of flow through porous media in the field of geotechnical engineering have been mainly in the saturated realm. Application of Darcy's law is widespread in these investigations, which include the consolidation of clays, and seepage through soils and rocks with heterogeneous and anisotropic conditions. Muskat (1937), Taylor (1948), and Leonards (1962) review the role played by Darcy's law in these investigations.

4.1.3. Hydrogeology

Groundwater movement is governed by Darcy's law. Freeze and Witherspoon (1966) developed a mathematical model to describe the two-dimensional flow pattern for a heterogeneous groundwater basin with any water table configuration. The authors utilized finite difference techniques to solve the equations of their model and examined the effects of changes in the water table configuration and geological formations on the flow pattern. The boundary conditions for the solution of the equation were that the groundwater basin was bounded on the bottom by an impermeable layer, on the top by the water table,

and on all sides by imaginary boundaries that simulated the groundwater divides. Freeze (1969) attempted to couple the saturated and unsaturated flow processes. He suggested that the unsaturated flow processes of infiltration and evaporation were in physical and mathematical continuity with the parallel processes in the saturated realm of recharge and discharge.

4.2 Validity of Darcy's Law

All of the disciplines of study just reviewed utilize Darcy's law for the quantification of water movement through porous media. However, an examination of the range of validity of Darcy's law should be made before it is applied to flow in forest soils. Darcy's law states that the relationship between the velocity of fluid flow and the hydraulic gradient is a linear one. The validity of this law requires that the flow be laminar. For larger pore sizes and flow velocities, flow can become turbulent. For such flows the velocity - hydraulic gradient relationship becomes non-linear and Darcy's law does not apply. The Reynolds number of flow is used as a criterion for deciding whether flow is laminar or turbulent. In straight tubes, the critical value of Reynolds number is in the range of 1000 - 2200. However, the critical Reynolds number at which the flow becomes turbulent is reduced when the path is tortuous. For flow in porous media, it is generally assumed that the velocity remains linearly related to the hydraulic gradient only if the Reynolds number is smaller than ten. Klute (1965) stated that laminar flow prevails in finer materials for any commonly occurring

hydraulic gradients found in nature. Only in coarse sands and gravels would nonlaminar flow conditions exist.

However, another consideration seems to be important in the surface layer of forest soils. Several researchers have expressed doubt about the formerly widespread belief that it is the bulk soil matrix that is serving as the primary conductive medium in undisturbed forest soils. Beasley (1976) comments that the response of subsurface flow to rainfall cannot be explained solely on the basis of saturated flow through the main soil mass. A comparison of the rather low saturated hydraulic conductivities of the bulk matrices of various textures of soil with those required to explain rapid response of interflow, has led researchers to speculate on other means by which water could move through the soil. As early as 1941, Hursh and Hoover postulated that the decay of roots and the channelling by microorganisms and small insects created relatively large continuous openings that could serve as hydraulic pathways for the rapid movement of water. Hursh and Fletcher (1942) observed that gravitational water must be transmitted at a rate much faster than had been originally considered to occur. Gaiser (1952) studied vertical and lateral channels formed by decayed root systems and asserted that the root channels could become pathways for the rapid movement of free water because they contain materials more permeable to water than the surrounding soil mass. Whipkey (1969) concluded from observations on a field soil pit that lateral flow was not moving as a mass interflow through the general soil matrix but rather through cracks and

channels in layered fine-textured soils.

However, the mere presence of cracks or channels does not ensure their conductive ability. Two important physical criteria must be met before these cracks and channels can become water conducting pathways. The first is that the water surrounding the crack or channel must be at a positive pressure before it will enter the crack or channel. One way this can be accomplished is for the channels to be open to the atmosphere with a positive head of water existing at the openings. Beasley (1976) stated that depressions, such as those formed by the uprooted trees or decayed stumps provide ideal locations for water to concentrate. He also suggested that decayed roots which may radiate outwards from the depressions provide natural pathways for water to concentrate. The pressure must remain positive all along the flowpath or else the water will move out of this pathway. The second criterion for the successful establishment of this type of subsurface flow is that these pathways, which are probably decayed root channels emanating from different sources, must all be interconnected. If at any location the continuity is broken, then the hydraulic conductivity of the soil matrix will become the limiting factor in the rate of flow. Whipkey (1969) seemed rather confident, based on his field observations, that the inter-connection between macro-channels and cracks, at least in the forest soil he observed, did exist.

These observations raise several questions. One pertains to the validity of hydraulic conductivity values obtained by laboratory measurements. Considering the importance of the

interconnection of the root channels, it is doubtful that laboratory measurements, even on undisturbed samples, would offer much more than an estimate of the possible field value of hydraulic conductivity. In this regard, use of the term hydraulic conductivity can be confusing. Since the term can often be interpreted to be a measure of flow primarily through the bulk soil matrix, a more general alternate term, conductivity coefficient is suggested as the term to be used to represent the transmission rate of soils containing cracks or channels. This term would include all other possible sources of water conduction besides just the soil pore space.

The combination of subsurface water conducting channels and high hydraulic gradients, which could be expected on mountainous watersheds, would most likely foster very rapid flow through the soil. Turbulent flow appears possible and has been observed at open soil faces (Whipkey, 1967). Thus, use of Darcy's law to quantify the flow of water in these instances might be questionable.

4.3 Extensions of Darcy's Law

Some researchers have viewed Darcy's law as being a specific case of a more general equation which states that the flux is proportional to the hydraulic gradient raised to a power. For Darcian flow, the exponent would be unity. Unfortunately the magnitude of this exponent varies unexplainably with variations in the flow regime, dependent on the diameter of the particles of the medium.

Several researchers have attempted to derive equations to quantify turbulent flow in porous media. Muskat (1949) suggested the addition of a second order term and his proposed equation took the form:

$$\frac{dp}{dl} = av + bv^2$$

with $\frac{dp}{dl}$ = pressure drop per unit length,
v = macroscopic velocity,
a,b = constants of the fluid and porous medium.

Ward (1964) applied the equation to six different kinds of porous media which included glass beads, sand and gravel and specified approximate values of Reynold's number for separating laminar and turbulent flow regimes. He presented the following equation of flow claiming it was valid for both laminar and turbulent flow in porous media:

$$\frac{dp}{dl} = \frac{\mu v}{k} + \frac{0.55\rho v^2}{k}$$

with p, l and v as in Muskat's equation,
and μ = absolute viscosity of the fluid,
k = permeability of the porous medium,
ρ = density of the fluid.

Unfortunately, problems would arise in the application of this equation to field conditions. Soil temperature can be extremely variable, due to soil heterogeneity, and thus effects on viscosity and in turn hydraulic conductivity would be significant. As a result, the application of this equation to the apparently turbulent flow in forest soils would not provide any more certain calculations of fluid flow in a field situation than would the

application of Darcy's law, which has already been shown to be widely applied in several different study disciplines and certainly offers the advantage of ease of application.

4.4 Flow in Layered Soils

The discussion on flow through porous media to this point has concentrated on homogeneous porous media. Most commonly, soil profiles are characterized by horizons having varying degrees of expression. Differences between horizons are often reflected in the water flow patterns through the soil profile.

4.4.1 Flow at the Boundary Between Two Soil Layers

Consider a soil having two distinct, isotropic layers with hydraulic conductivities K_1 and K_2 with layer 1 overlying and being more permeable than layer 2. At every point along the interface of the two layers, since the suction heads and elevation heads are equal on either side of the boundary, the hydraulic heads and hydraulic gradients must also be equal. Let t represent the direction tangential to the interface and n represent the direction normal to the interface. Let the flow make an angle B_1 with the normal in layer 1 and angle B_2 in layer 2. The requirement for continuity in the direction normal to the interface, at the interface, suggests that, from Darcy's law:

$$K_2 i_{n2} = K_1 i_{n1}$$

where i represents the hydraulic gradient. Also since:

$i_{t1}/i_{n1} = \tan B_1$ and $i_{t2}/i_{n2} = \tan B_2$,

then $\tan B_2/\tan B_1 = K_2/K_1$

Therefore, upon passing from a pervious layer to a less pervious layer, the flow streamlines intersecting the boundary are refracted towards the normal, with the refraction becoming more pronounced the larger K_1 is relative to K_2. As B_1 approaches 90^o, the refracted streamline within the less pervious layer shows a greater and greater lateral component (Bear et al., 1968).

Zaslavsky (1964) noted that when flow is almost horizontal (parallel to the layers), a ten-fold reduction in the hydraulic conductivity from one layer to the next may be considered as the formation of an impermeable layer. If large hydraulic gradients exist, in a downslope direction, lateral flow could become significant under these conditions.

4.4.2 Flow in Layered Isotropic Soils

4.4.2.1 Flow Parallel to the Layers

Consider a three-layered soil profile comprised of three layers of thickness d_1, d_2, and d_3 respectively and having hydraulic conductivities of K_1, K_2 and K_3 respectively (Figure 5a). For flow parallel to the layers, the loss of head H along a common length L and the hydraulic gradient H/L are the same for all layers. The total rate of flow is then given by:

$$Q = (K_1d_1 + K_2d_2 + K_3d_3)\ H/L$$

The same flux Q would be produced in a homogeneous soil of the

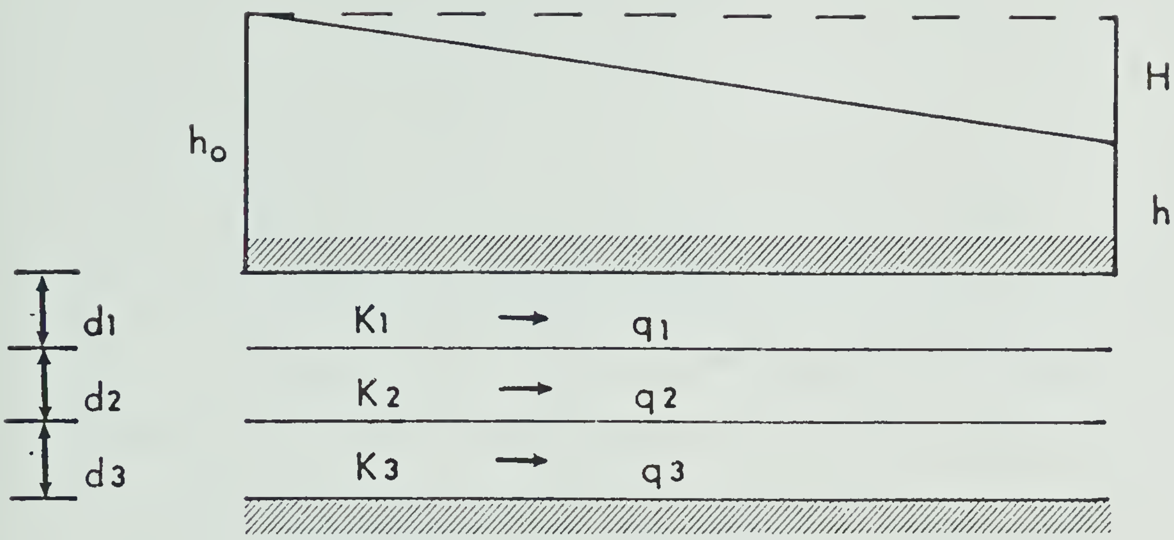

Figure 5a Flow parallel to layers (Bear *et al.*, 1968)

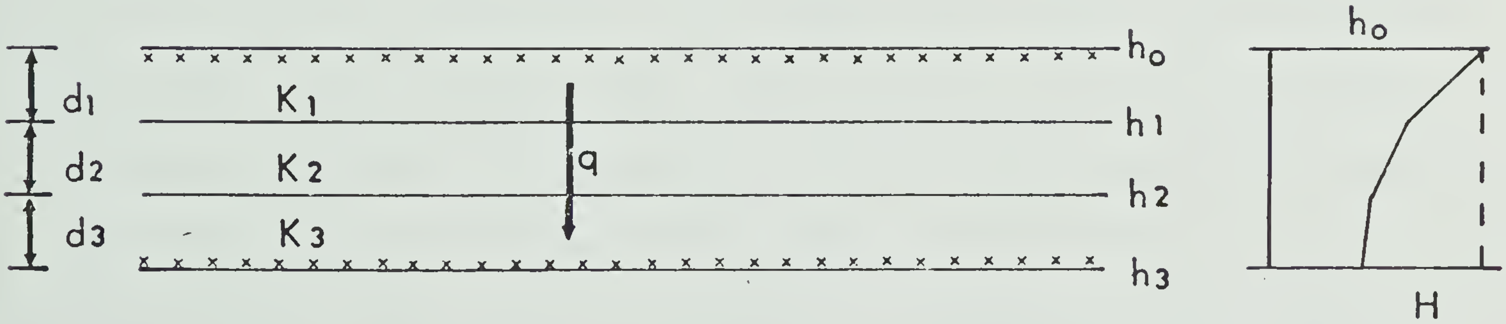

Figure 5b Flow normal to layers (Bear *et al.*, 1968)

same total thickness and hydraulic gradient if it had an hydraulic conductivity of:

$$K' = \frac{K_1 d_1 + K_2 d_2 + K_3 d_3}{d_1 + d_2 + d_3}$$

(Bear et al., 1968).

This equation can be used to show that one highly pervious layer renders the entire system pervious parallel to the layers.

Selim et al. (1975) presented an analytical solution for a two-dimensional multi-layered hillside seepage problem. The flow medium is assumed to be water saturated to the soil surface, bounded below by an impermeable barrier at infinite depth. The authors suggest that the soil geometry would be that of seepage from a sloping soil to a creek. Results worthy of note here are those pertaining to the effect of hydraulic conductivity variation on the flow pattern. For the two-layered case examined, with the upper hydraulic conductivity being 10 times that of the lower, 90% of the flow occurred in the upper layer. In a three-layered soil with the upper and lower layers having hydraulic conductivity values 10 times that of the middle layer, the upper and lower layers accounted for over 70% and 20% of the flow respectively. Both results show that a permeable layer near the surface will carry the bulk of the flow, when underlain by a less permeable layer.

4.4.2.2 Flow Normal to the Layers

Consider the same three layered soil profile but now with flow in the normal direction (Figure 5b). The flux through each

layer has a different partial head loss and a different hydraulic gradient. The resistance to flow of a single layer can be specified as $r_i = d_i/K_i$. For the three layer system the total resistance is given by:

$$R = d_1/K_1 + d_2/K_2 + d_3/K_3$$

The specific flux q is given by:

$$q = H/R = H/(d_1/K_1 + d_2/K_2 + d_3/K_3)$$

The same specific flux could be produced in a homogeneous soil under the same conditions if it had an equivalent hydraulic conductivity of:

$$K'' = \frac{d_1 + d_2 + d_3}{d_1/K_1 + d_2/K_2 + d_3/K_3}$$

(Bear *et al.*, 1968). This equation can be used to show that one impervious layer renders the entire flow system impervious in the normal direction.

Comparison of the equations for K' and K'', for given values of K and d, will show that the equivalent hydraulic conductivity parallel to the stratification is always greater than the equivalent hydraulic conductivity perpendicular to the stratification.

4.5 The Effects of Anisotropy on Flow

The preceding discussion on flow in layered soils considered the nonhomogeneous effects of soil layering, showing that a layered soil system can be replaced by an equivalent homogeneous one. In addition, often the hydraulic conductivity of a soil at a point has different values in different directions due

to sedimentation, cracks, rootholes or a preferentially oriented arrangement of soil particles. Such soils are said to be anisotropic with respect to hydraulic conductivity. When the flow problem involves components of flow along more than one direction, anisotropy can be taken into account by a transformation of the coordinates in such a way that the system becomes an isotropic medium. Maasland (1957) showed that this equivalent homogeneous isotropic medium has an hydraulic conductivity given by:

$$K = (K_x K_y K_z / K_o)^{1/2}$$

where K_o is an arbitrary constant and K_x, K_y and K_z are the hydraulic conductivities along the principal directions of the anisotropic medium.

Infiltration is usually assumed to be vertical unsaturated flow into the soil. However, the actual direction of this flow into a soil slope is highly variable, depending upon the degree of anisotropy of the soil layers, the slope itself and changes in shape along it. The hydraulic conductivity parallel to the soil horizons is always larger than the conductivity normal to these horizons (Bear _et al._, 1968). The effect of anisotropy is that flow does not necessarily occur in the direction of the maximum hydraulic gradient. In an anisotropic soil, where the horizontal hydraulic conductivity generally exceeds the vertical hydraulic conductivity, the direction of flow will be along a line closer to the horizontal axis. If a force deviates to one side of the direction normal to the soil layers, the flux vector will deviate even further away from the normal on the same side. In a

sloping soil, the combined force of gravity and pressure (hydraulic head) will usually deviate away from the normal in a downslope direction. Therefore, under these conditions, one would always expect, in addition to the flow component normal to the soil horizons, a lateral flow component parallel to the soil layers in a downslope direction. The relative magnitude of the lateral flow component and the normal flow component depends on the degree of anisotropy of the soil, that is, the ratio of the hydraulic conductivity in the lateral direction to that in the normal direction.

4.6 The Effects of Slope on Interflow

If the soil is homogeneous, the effects of flow anisotropy will be more pronounced on the steeper slopes. The flow direction of the infiltrating water will be diverted further and further downslope as the steepness of the slope increases. Convergence or divergence of streamlines will occur as a result of the interactions of anisotropy and slope shape changes. When slopes are almost level, streamlines converge slowly to a point deep beneath the surface. As curvature increases and slopes steepen, streamline convergence occurs at shallower depths. In general, once there is some profile differentiation, the infiltration streamlines will tend to diverge on the convex portions of the landscape and to converge on the concave portions of the landscape.

The above discussion has shown that the anisotropic properties of a soil profile as well as slope shape can be

instrumental in the establishment of a lateral flow component to infiltrating water. However, these factors in themselves would not create lateral flow of any great significance. Slope gradient provides the chief driving force for lateral flow, primarily determining the quantity of interflow that will prevail, with steeper gradients causing larger flows. Forested watersheds located in the headwaters of most rivers often exhibit such steep gradients. As they also usually have some sort of vertical flow impediment, an excellent opportunity for lateral flow in these watersheds exists. In nature the situation is further complicated by the fact that soils tend to vary consistently with slope gradient within a catchment. The most usual relationship is for soils to be more permeable on steep slopes, so that subsurface flow velocities are increased in them. On gentle slopes in humid regions slow drainage may lead to the development of peaty A horizons, which can hold large quantities of water and also impede lateral flow because of their low hydraulic conductivities. Soils also tend to be thicker on gentle slopes, thereby increasing their water holding capacity and thus partially counteracting their tendency to saturation. On a concave slope, the converging of interflow causes the saturated layer to become thicker, and this may extend saturated conditions to the surface. Hollows are examples of such concave slopes and tend to be the most dynamic in their response to storm precipitation. Slope convexity permits the saturated flow velocity to increase downslope due to increasing slope gradient. In this case, the downslope discharge will increase without an increase in saturation depth which

could bring the saturated layer to the soil surface.

4.7 Other Factors Affecting Interflow

Vegetation cover directly affects the maintenance of infiltration capacity and the conditioning effect of organic matter on soil structure and porosity. The most noteworthy effects of cover are found in undisturbed forest stands (Gaiser, 1952; Chamberlin, 1972). Under these conditions decaying root systems create important channels for free water conduction. Fine textured or well-layered soils have their saturated hydraulic conductivities increased by the presence of these channels as well.

Land use, highly interrelated with vegetative cover, commonly has the greatest effects on infiltration and, as a result, affects the amounts of precipitation occurring as overland flow and subsurface flow.

Climate acts directly through rainfall intensity and in conjunction with vegetation on the rates of evapotranspiration to determine the antecedent conditions of the watershed through the soil moisture regime. Where deep soils have soil moisture thoroughly depleted by dense vegetative cover during the growing season, interflow will not occur until soil moisture defecits are satisfied through moisture infiltration. These defecits will be a maximum when evapotranspiration is at a maximum.

CHAPTER 5

ASSESSMENT OF THE ROLE OF INTERFLOW

An assessment of the role of interflow in any given watershed can proceed in one of two ways: (1) A technique using hydrograph analysis whereby a hydrograph is used as input and inferences are drawn about the flow processes from its analysis, or (2) a modelling technique whereby the precipitation is used as the input and the outflow is simulated. Modelling represents the hydrologic cycle, or some portion of it, and agreement between the simulated and actual outputs implies that the modelled flow processes may have been correctly described.

5.1 Hydrograph Analysis

Use of this technique implies that given the discharge hydrograph, which is some combination of the various flow processes, one can indeed split the hydrograph in a quantitative fashion into the respective flow processes comprising it. Generally the hydrograph is divided into two components: baseflow and direct runoff, which includes overland flow and interflow. Butler (1957) states that subsurface flow may be arbitrarily included in the direct surface flow and baseflow or estimated separately from experience. Chow (1964) takes an intermediate position, dividing interflow into two parts: prompt interflow which is added to direct runoff, and delayed interflow which is added to baseflow. This division is qualitative and a

quantitative interpretation is rather difficult. Unfortunately this lumping technique suggests that overland flow and interflow are very similar flow processes and can be combined. As a result, this lumping technique offers little information about interflow.

Barnes (1939) suggested a technique for separating the hydrograph into three components. This technique appears in most standard texts on hydrology and is as follows:

(1) Plot the hydrograph on semilogarithmic paper.

(2) Approximate the groundwater recession by a straight line extended back under the hydrograph.

(3) Plot the residuals (surface runoff and interflow).

(4) Fit a straight line to the recession of this curve and extend under the hydrograph.

(5) Plot the residuals as surface runoff.

In this analysis, the rising limb of the groundwater and interflow hydrographs must be approximated. Step (2) has the inherent assumption that the values considered on the recession limb are those taken after a sufficiently long time so that storage from interflow and surface runoff have been released. Problems can arise because neither the position of the peak of each flow component nor the shape of the rising limb of the discharge hydrograph is well defined. Chow (1964) assumed, quite arbitrarily, that the peaks of interflow and baseflow fall under the inflection point of the recession hydrograph.

Success with Barnes' method for hydrograph analysis is not widely reported. In fact the opposite is true. Linsley and Ackermann (1942) claim they were unable to identify the interflow

component using the technique. Kulandaiswamy and Seetharaman (1969) were extremely critical of the technique.

Thus a simple, yet reasonably accurate hydrograph separation technique is still unavailable. Judging by the comments of various researchers, development of such a technique may never come. Snyder (1955) and Freeze (1973) were both rather critical of the continuing use of hydrograph separation and imply that the exercise is a fruitless one. Nemec[2] (1972) summed up the technique of hydrograph separation quite succintly: "...the selection of the method (of hydrograph separation) is quite arbitrary. Simple methods, are, however, recommended since the precision of the separation is very dubious by any method."

5.2 Hydrologic Modelling

One must therefore attempt to gain information about the flow processes that created the hydrograph. Research for this purpose has been of two types: field measurement in representative experimental drainage basins and theoretical studies using mathematical hydrologic models. Most of the field research has been based on instrumentation in units much smaller than full watersheds, often on individual slopes that feed short reaches of small tributary streams. In evaluating effects of land use, one requires an overall view of the basin rather than a small portion. Thus hydrologic modelling of the

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[2] J. Nemec. 1972. Engineering hydrology. McGraw Hill. p. 240.

entire basin must be attempted. Concentration in this study will be made on small watersheds where the land flow processes are more dominant than the channel flow. In addition, these watersheds are probably the ones most likely to have the shape of their hydrographs significanty affected by land use changes. This emphasis would also correspond to the concentration on the area surrounding a first order stream in large watersheds. Hewlett (1974) comments that no one has yet provided a physically valid model for the first order stream.

In this technique of before-the-fact analysis, a model of the watershed is chosen or built with all relevant physical parameters included. The model is 'tuned' until the simulated hydrograph approximates the actual one. In this way information can be gained about the physical flow processes that are contributing to the hydrograph of the watershed in question.

However, before a decision can be made as to whether an appropriate model can be chosen or must be built, background information about the philosophy of modelling and the criteria for model selection should be examined.

5.2.1 Choosing a Model

Many hydrologic models are available for use today. They can be categorized as two main types: stochastic or deterministic. Stochastic models are those which, using the statistical properties of existing records and probability laws, generate future hydrologic events, mainly because in many cases hydrologic data for extreme events may be lacking. Deterministic

models attempt to specify the hydrologic processes through mathematical functions. In deterministic models, when the input data, boundary conditions and initial conditions are specified, the output is known with certainty. Discussion will hereafter be limited to deterministic models. Models can be broken down into two further classes: conceptual and physical. Conceptual deterministic models treat the actual physical processes in generally a superficial way. This is in sharp contrast to pure physical deterministic models where parameters are selected that have a physical meaning and can be physically measured or estimated from physical data.

A watershed manager would like to predict what effect changing some watershed parameter will have on watershed output. Theoretically this can be done with physical deterministic models since the parameters are related to physical quantities, but not so easily for conceptual models since there is no way of relating the change in watershed characteristics to the parameters of the model. Conceptual models which have their parameters optimized using a calibration period are particularly susceptible to this problem. Despite the advantage of physical deterministic models in this regard, few truly physical models exist, mainly due to the limited knowledge scientists have acquired about individual processes that take place in the watershed. Empiricism can be found in the quantification of many of the processes and empiricism can only imply conceptualism.

The final categorization of models separates them into

lumped or distributed models. In lumped parameter models precipitation, topographic features, vegetation and soils are all considered to be homogeneous over the watershed. There are, however, very few instances in nature where this would hold true. For example, the spatial variation of soils is generally the rule in nature. Distributed models attempt to incorporate these spatial variations, retaining the positional uniqueness of them. Obviously, these models would be most valuable in predicting the effects of vegetation manipulations on water yield, since, for example, partial clearcutting of vegetation can only be handled by a distributed model.

Distributed models also possess the inherent ability to model conditions at several (all) points within the watershed simultaneously, offering the freedom of multiple checks upon the results if several sub-basins exist within the watershed. These models also offer the opportunity to incorporate relationships developed on small scale plot-size studies because of the smaller subareas considered. This incorporation would not require the use of a 'scaling factor' which would have to be used for lumped models if results from small plots were to be extrapolated to much larger, and most likely, more nonhomogeneous areas.

The primary disadvantage of a distributed model as compared to a lumped model is the additional burden of computational requirements that tend to increase as catchment size increases. With the advent of more sophisticated computers, this added burden need not be a severe one. Only through the use of distributed models can hydrologists hope to account for

the many intricacies of as complex a system as the hydrologic cycle.

Huggins et al. (1973) point out that the potential for accurate modelling of any catchment is far greater with a distributed model than with a lumped model because of the flexibility for considering an almost unlimited range of spatially varying conditions. This potential is only realized if the model is designed to take advantage of all the additional information that can be offered in a characterization of the watershed. The basic premise of lumped models is that, in spite of the recognized spatial variability of these factors, they have a non-significant effect on the water yield. On the other hand, distributed models are designed to account for what is believed to be significant variability.

5.2.2 Review of Existing Distributed Hydrologic Models

A review of the types of existing hydrologic models reveals that few of them are distributed models (Fleming, 1975). Two models considered as being distributed are those of Huggins and Monke and Schultz. These models will be reviewed, as will the Kozak model, a partially distributed model.

5.2.2.1 Huggins and Monke

The Huggins and Monke model (Huggins and Monke, 1967) is a specific purpose model concerned primarily with surface runoff. A flow chart for the model is presented in Figure 6. This model

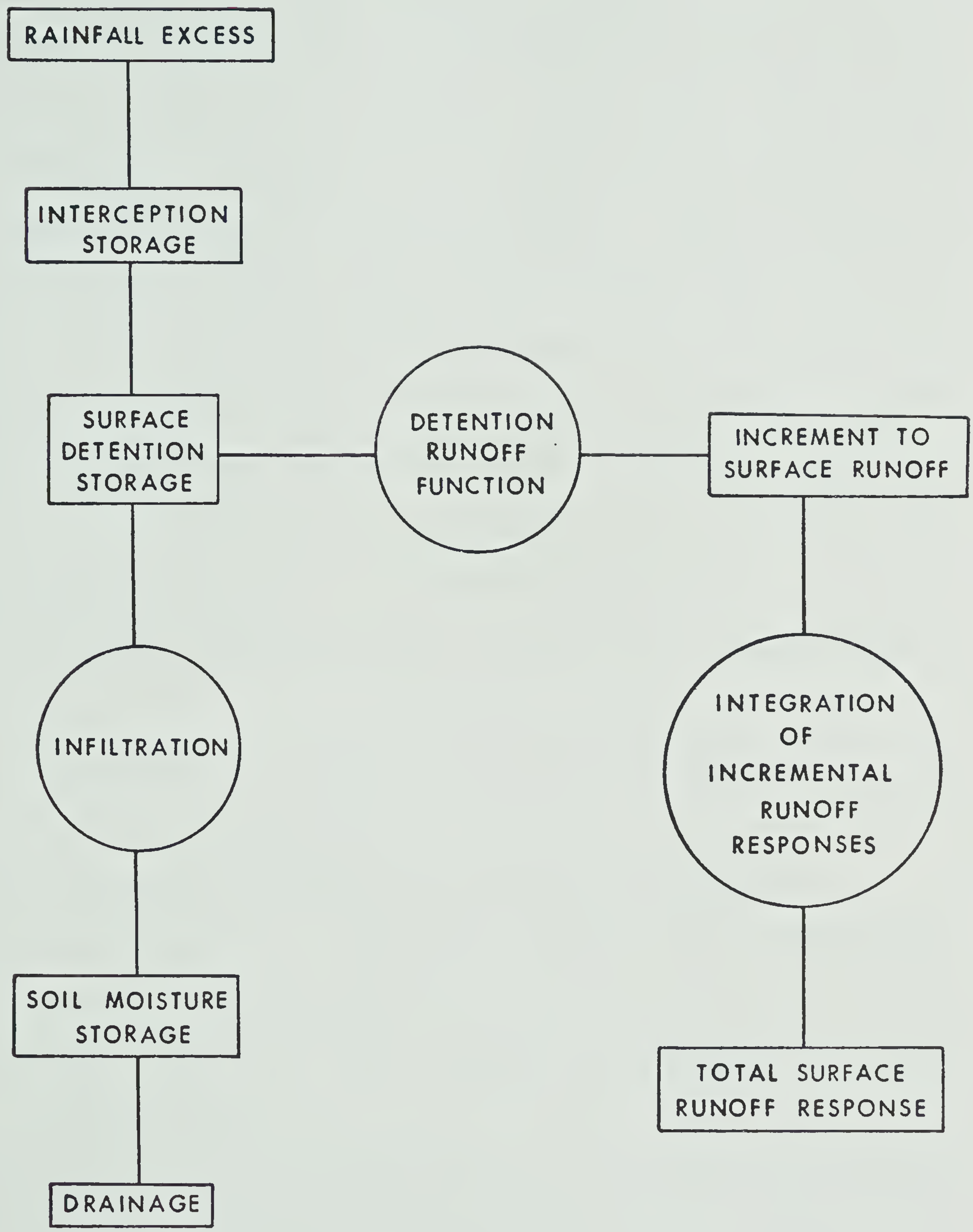


Figure 6 Huggins and Monke model structure (Fleming, 1975)

is not a complete representation of the hydrologic cycle since it neglects the subsurface components of flow and evapotranspiration. The most noteworthy feature of the model is the concept of finite elements to represent land surface features. An element is defined by the authors as an area within which all hydrological parameters are uniform. The catchment is subdivided into a grid of square elements, each of which is specified with a surface slope, slope direction and soil type number. The response of each element is described by deterministic equations which characterize infiltrations, surface detention, interception and surface runoff. The entire watershed response to a given storm is analyzed by integrating the continuity of mass equation over the entire catchment.

The primary inadequacies of the model are (Huggins et al., 1976):

(1) Inability to specify non-uniform vegetal cover and hydraulic roughness,

(2) No means of designating channel flow roughness,

(3) Assumption of negligible interflow and groundwater contribution to runoff,

(4) No provision to handle non-uniform rainfall distributions.

As the authors suggest, inadequacies (1) and (2) can be fairly easily removed. Inadequacy (1) would have to be removed before vegetation manipulation simulations could be attempted while inadequacy (2) would have to be removed for applications to large watersheds where channel flow would play a progressively

more important role. The assumption specified in inadequacy (3) can be limiting, particularly in applications to forested watersheds where interflow dominates the flow process. In the model, evapotranspiration is not calculated and thus one may question the validity of the soil moisture storage values, which are important in the calculation of infiltration. This is particularly critical since the model utilizes the Holtan equation for infiltration, which relates the amount of infiltration to the available storage remaining for infiltrating water. However, the model could still give 'reasonable' results for surface runoff because an unusually high drainage contribution to groundwater would result in soil moisture storage values comparable to those caused by more realistic lower values of groundwater recharge coupled with evapotranspiration losses. Since infiltration is essentially regarded as a loss (water becomes unavailable for surface runoff) a more realistic apportionment of infiltrating water may be unnecessary if the model is used specifically for surface runoff prediction.

Huggins et al. (1973), using two watersheds having areas of 33 and 18 ha in size, found the model to be particularly sensitive to changes in the values of the roughness coefficient (affected the runoff peak), antecedent soil moisture and surface retention depth coefficient (affected the volume and the first of double peaks in the flow). The interception parameter for high runoff producing storms became a relatively minor contributor but the underestimation of the recession limb of the hydrograph suggested that perhaps interflow was not a negligible component

in the two watersheds.

5.2.2.2 The Schultz Model

Schultz (1968) utilized a linear distributed system for flood hydrograph synthesis in Germany. The model (HYREUN) utilized as input information rainfall data according to a temporal and spatial distribution. For application of the HYREUN model (see Figure 7 for its flowchart) isochrones, lines of equal travel times, were first constructed by determining the time of concentration for the individual reaches using the Soil Conservation Service nomogram which were then divided into convenient equal time intervals (1/2 hour increments for medium sized catchments). Corresponding to the choice of the isochrone time intervals, the catchment was subdivided into a grid of elemental areas, in proportion to the magnitude of the time elements. To each element of area was assigned a lag determined by the isochrones.

The time distribution graph of total precipitation was used to derive the time distribution of precipitation within the elemental areas with the time pattern assumed to apply within the Thiessen polygon formed around each recording rain gauge. The model was designed specifically for flood hydrograph synthesis and is incomplete as a land use model since it does not include ground water flow, interflow or evaporation.

5.2.2.3 The Kozak Model

The Kozak Model (see Figure 8 for the flowchart) can be

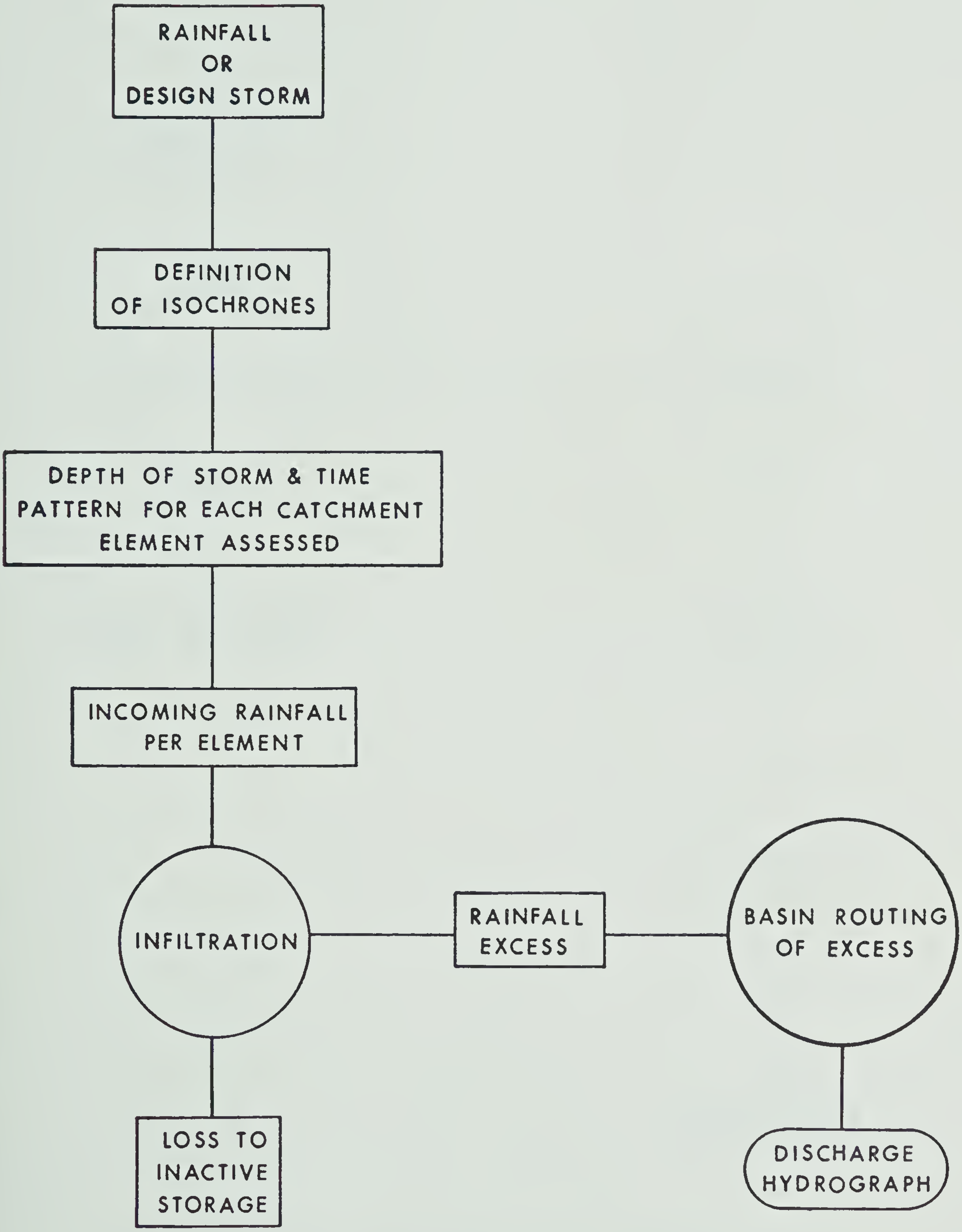


Figure 7 HYREUN model structure (Fleming, 1975)

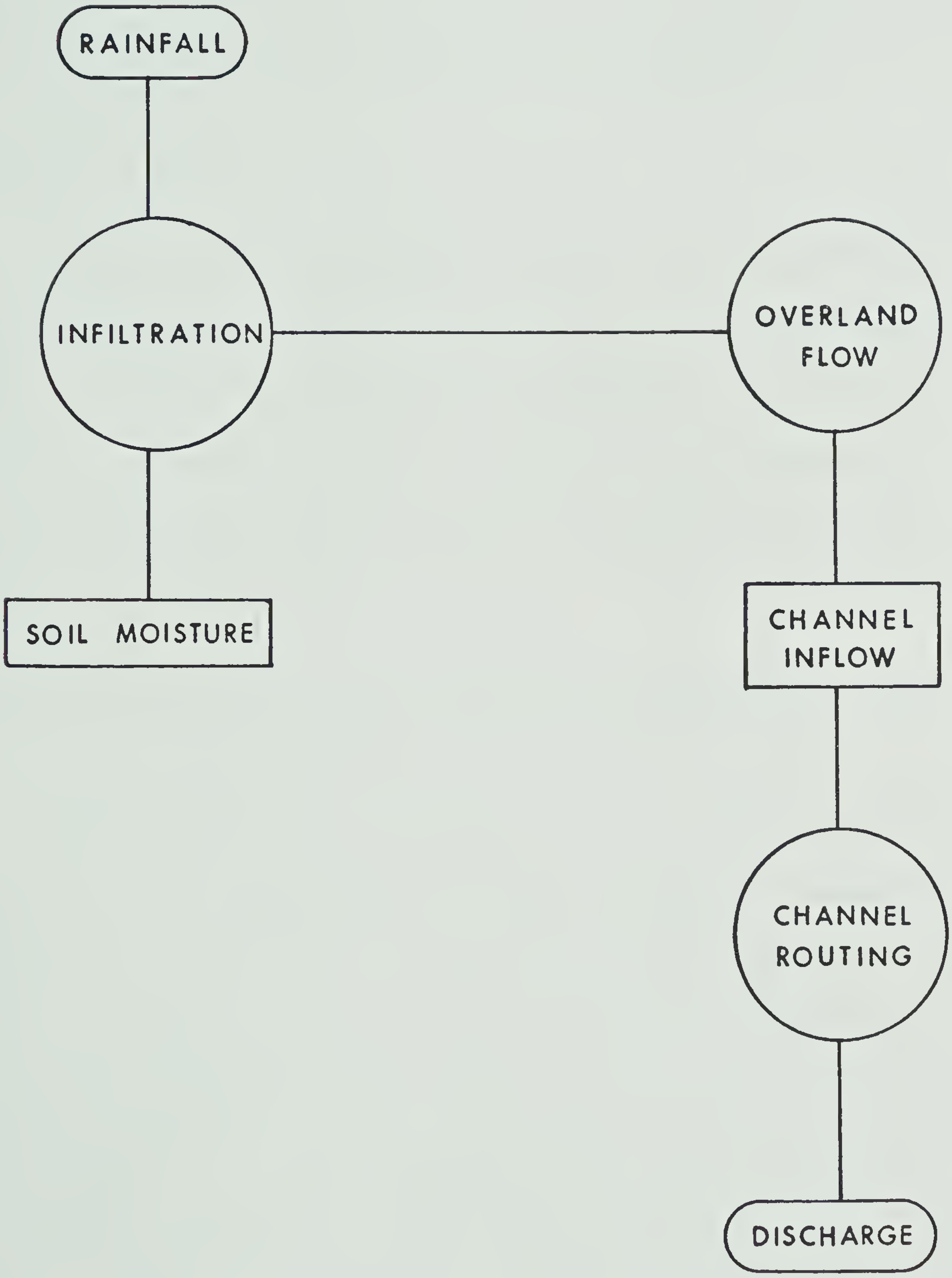


Figure 8 Kozak model structure (Fleming, 1975)

considered to be partially distributed because the catchment surface is subdivided into a large number of homogeneous units, each allocated ten parameters defining its characteristics (Kozak, 1968). Since the model represents basically only the overland flow and channel flow components of the hydrologic cycle, most of the parameters are related to these processes.

5.2.2.4 Summary

These three hydrologic models, reviewed because of their potential appropriateness due to their distributed nature, can be seen to be process specific (generally overland flow oriented) and are actually incomplete models of the hydrologic cycle. Thus they would be of little value in attempts at modelling subsurface flow.

5.2.3 Examination of the treatment of subsurface flow by currently used hydrologic models

Several hydrologic models are currently available that are aimed at the evaluation of the hydrologic impacts of forestry activities. Three such models are the Subalpine Water Balance Model (WATBAL), a snowmelt model; the PROSPER model, an evapotranspiration model; and the Leaf and Brink Model which concentrates on snow accumulation and snowmelt in subalpine watersheds (Troendle, 1979; Leaf and Brink, 1973). Unfortunately none of these includes any subsurface flow routing. Thus these models will be eliminated from further consideration.

There are other models available which are general purpose hydrologic models but yet are not distributed. They are worthy

of examination here because they do make an attempt to include subsurface flow. Three such hydrologic models will be briefly reviewed in an attempt to analyze their quantitative treatment of subsurface flow. The three models to be examined are the SSARR model, the Stanford model, both lumped models, and the USDAHL model which concentrates mainly on land flow processes (smaller watersheds). The Stanford model can theoretically be used for both large and small watersheds. The SSARR model and the Stanford models are reviewed because of their universal use and the USDAHL model because of its physically correct depiction of the soil hydrologic properties.

5.2.3.1 The SSARR Model

The SSARR model essentially divides rainfall into input available for surface, subsurface or baseflow runoff and computes these with empirical relationships between runoff percentage (ROP) and the soil moisture index (SMI) which is a measure of relative soil wetness, being zero at the wilting point and at its maximum value at field capacity. The total generated runoff (RGP) for a given period is calculated by:

$$RGP = ROP \times WP$$

with ROP = runoff percentage, and

WP = weighted precipitation for the period.

A SMI value is calculated at the end of each period by:

$$SMI_2 = SMI_1 + (WP - RGP) - (PH/24 \times KE \times ETI)$$

with SMI_1 = SMI at the beginning of the period,

SMI_2 = SMI at the end of the period,

PH = period length,

ETI = evapotranspiration index,

KE = factor reducing ETI on rainy days.

When no precipitation occurs, KE = 1.0 and SMI will be reduced by the constant factor PH/24 X ETI. SMI will increase when (WP - RGP), which represents rainfall not contributing to runoff, exceeds the adjusted evapotranspiration factor.

A portion of runoff excess contributes to the baseflow component (BFP) and is used as input as a function of the baseflow infiltration index (BII) with a prescribed relationship (table of values). BII is computed for each period according to an equation including RG, the runoff rate (RG = RGP/PH). Knowing BII, the specified table is entered to get the appropriate value of BFP, which when multiplied by RGP specifies the amount of baseflow. The input to surface and subsurface runoff (RGS) is then computed by:

$$RGS = RG \times (1 - BFP)$$

with RG defined as above. Residual total runoff is divided into surface runoff and subsurface runoff through the use of another table specified for the particular basin, giving surface runoff as a function of RGS. Subsurface flow can then be calculated by subtraction. Each of the computed component inflows are then routed through a specified number of increments of reservoir type storages to obtain streamflow (U.S. Army Corps of Engineers, 1971).

5.2.3.2 The Stanford Model

Crawford and Linsley (1966) consider two soil zones in their Stanford Model. The upper zone can be considered to represent the upper few centimeters of soil which react quickly to rainfall and control overland flow, while the lower zone represents the soil moisture storage capacity. When rainfall occurs, moisture is divided between surface detention and infiltration based on Holtan's equation. Percolation from the upper zone is calculated on the basis of the following equation:

$$D_r = 0.1 \times INF \times UZSN \times (UZS/UZSN \times LZS/LZSN)$$

with D_r = drainage from the upper zone,

INF = parameter,

UZS = actual upper zone storage,

UZSN = normal upper zone storage,

LZS = actual lower zone storage, and

LZSN = normal lower zone storage.

The amount of water entering the groundwater zone from the lower zone is calculated using one of a series of equations, dependent upon the ratio of LZS/LZSN, which forms the basis for the equation.

The water available for interflow is calculated by the equation:

$$I_{INT} = K \times 2^{LZS/LZSN}$$

where K is a coefficient. Water from interflow storage is assumed to enter the stream channel at a rate based on the recession rate obtained from an analysis of recorded streamflow hydrographs:

$$Q_i = (1.0 - (IRC)^{1/96}) \times SRGX$$

with Q_i = interflow volume,

IRC = daily recession rate of interflow, and

SRGX = volume of interflow storage.

5.2.3.3 The USDAHL Model

In this model, soils on each watershed are grouped according to land capability classes to form hydrologic response zones, typifying the physiographic sequence, such as uplands, hillsides, and bottom lands in these areas. Water that infiltrates is proportioned to evapotranspiration, to downward seepage or to lateral return flow in each flow regime. Channel flows and subsurface return flows are routed by the simultaneous solution of the continuity equation and a storage function. Storage coefficients are obtained by the evaluation of the flow recession curve for a given watershed. Flow from each unit is routed separately through watershed storage and then all are summed to obtain watershed outflow (Holtan *et al.*, 1975).

Several regimes of subsurface flow are delineated based on hydrograph separation. Increments of downward seepage to the next regime are computed as a function of the gravity water present:

$$Q = \Delta t \times C \times (G - SA)/G$$

Q = water passing downward to next regime,

Δt = time increment,

C = rate of downward seepage,

G = gravity water, and

SA = air space equivalent of water.

The potential rate of outflow from a regime is given by:

$$q_2 = \frac{2\,\Delta I}{2m + \Delta t} + q_1 \frac{2m - \Delta t}{2m + \Delta t}$$

with q = rate of outflow,

Δ I = inflow volume,

m = routing coefficient,

Δ t = time increment, and

1 and 2 = beginning and end of Δ t.

5.2.3.4 Summary

Distributed models have been shown to possess attributes that would facilitate the assessment of interflow and ultimately in the prediction of the effects of land use manipulations on streamflow. Unfortunately few current hydrologic models are distributed in nature and those that are, are exceedingly deficient in representing the hydrologic cycle. Furthermore, the treatment of interflow by three other well known models was shown to be rather empirical in nature and would thus offer little benefit in the consideration of interflow and watershed manipulation.

Evidence from experimental studies has already been presented showing that a saturated layer of water is formed within the soil due to a flow impeding, subsurface layer, and that due to large gradients, the water in this layer would flow laterally, forming interflow. A hydrologic model being developed for assessing interflow should be premised on these conditions. The simulation of the saturated layer and the formation of the

lateral flow component within the model should be physically based, taking into account the water holding and transmissive properties of the soil.

CHAPTER 6

DEVELOPMENT OF A HYDROLOGIC MODEL

A search of the literature has shown that existing hydrologic models cannot be suitably adapted to evaluate the hydrologic responses of watersheds to land use manipulations, with interflow being the dominant flow. This suggests that a new model must be developed to achieve this objective.

6.1 General Attributes

A hydrologic model designed to evaluate land use effects should have the following attributes:

(1) Realistic and representative: The model must be able to describe correctly the soil profile as a water reservoir in order that plant use and subsurface flow simulations can closely depict those actually occurring in nature.

(2) Flexible: The model should be able to characterize the spatial variation of watershed characteristics such as slope, vegetative cover, soil depth and water holding capacity, as well as be able to handle the spatial and temporal variation of precipitation.

(3) Simple: The model must be able to function with a minimum of required input, hopefully with data that are not difficult or costly to obtain.

(4) Sensitive: The model must be able to respond to and show changes in soil moisture (affected by plant use) as well as in subsurface flow in order that the effects of land use changes can be evaluated. The response must also occur within a range of parameter values that would be found in the field.

(5) Accurate: The model should produce results within an acceptable degree of accuracy so that some credance can be placed in the model when it is used for predicting effects of land use changes. Accuracy could be assessed by applying the model to a real-life situation with known outputs for given inputs, for example, comparing recorded streamflow values with simulated ones.

(6) Adaptable: The model must be easily adaptable to any type of watershed, or part thereof, for example, sub-basin or hillslope.

6.2 Specific Attributes

6.2.1 Type of Model

The model developed and discussed in this treatise offers a distributed feature which considers the spatial variations of watershed parameters by dividing the watershed into small, homogeneous units, hereafter called elements, of equal area through the use of a grid system. Since the complete set of watershed parameters can be varied from element to element, the model should be easily adaptable to any watershed. The model

developed herein will be called the SLUICES model (an acronym for Soils and Land Use affecting Interflow and Creating Effects on Streamflow).

The square element technique has been utilized by other researchers. Kouwen (1972) utilized a square element system utilizing an element size of 1 km X 1 km. Huggins et al. (1973) used the concept of finite elements of land surface in their model and applied it to two watersheds of size 18.4 ha and 33.2 ha using element sizes that varied from 45.7 to 91.5 m. Charbonneau et al. (1975) utilized a similar type of square element grid system. Their element size was 10 km X 10 km, corresponding to the UTM grid, and they applied their CEQUEAU model to watersheds of very flat slope.

6.2.2 Model Parameters and Operation

Because of the important role that the soil, being the conductive medium, plays in subsurface flow, importance will be attached to the appropriate modelling of the soil profile as a water reservoir. This will be accomplished through the use of the commonly accepted soil parameters of saturation, wilting point and field capacity, all of which refer to moisture content percentages. The effects of various soil types will be reflected in the changing values of these three parameters. Water between saturation and wilting point makes up total storage and is subject to evapotranspiration loss while water between saturation and field capacity is subject to drainage. The depth of the soil profile is another parameter with the conductive properties of the

soil profile being characterized by a conductivity coefficient.

Model operation hinges on the presence of horizons of different natural permeability in the soil profile causing infiltrating water to be delayed in its downward percolation forming a saturated layer of water at the top of a horizon with lower permeability. Formation of this saturated layer due to the presence of an impeding soil horizon has been well established through numerous field studies (Harr, 1977; Beasley, 1976; Weyman, 1973; Betson *et al.*, 1968; and Whipkey, 1965). Steep slopes, coupled with high soil conductivity coefficients, can cause the water to flow laterally, forming interflow.

Additions to the soil moisture reservoir can be in the form of precipitation as well as inputs from adjacent elements either as subsurface and/or overland flow. Depletions take the form of evapotranspiration, loss through subsurface flow to adjacent elements or to deep percolation. Should field capacity be exceeded due to a net gain of water, then the moisture in the profile is redistributed to give two distinct zones within the soil profile: a zone having a moisture content at field capacity and a saturated zone, forming at or above the less permeable layer. The depth of the saturated layer will be dependent upon the amount of water above field capacity which can be used to bring the moisture content of a portion of the soil profile up to saturation. In all cases, the depth of the saturated zone plus the depth of the unsaturated zone must equal the total profile depth. Should the net addition of water be large enough, it is possible for the complete profile to become saturated from below. In this case an overland flow component will form. After an

appreciable post-storm period, the soil profile will revert to being a completely unsaturated profile with an overall moisture content less than field capacity, at which time the model simulates a soil moisture budget.

Model operation is based on the delineation of three storages: overland flow storage, unsaturated flow storage, and saturated flow storage. However, all three storages cannot exist simultaneously. Existence of the unsaturated storage indicates that at least some portion of the soil profile is unsaturated. As soon as field capacity for the soil profile is exceeded, saturated flow storage is created, at the expense of part of the unsaturated storage, indicating that a portion of the soil profile is unsaturated and a portion is saturated. If just enough water is added to the profile to completely saturate it, then the profile will consist of only the saturated flow storage. If more water is added, overland flow will occur. If the supply of water is curtailed while evapotranspiration and outflow continues, overland flow storage will be depleted first. At some point enough water will have been depleted from the soil profile so that unsaturated flow storage will be created and will co-exist with the saturated flow storage. As evapotranspiration and subsurface flow continue with no additions of water, the saturated storage will be depleted and only the unsaturated flow storage will be left, depicting an unsaturated soil profile again.

Beside the evapotranspiration output, each of the model storages can have both a lateral and vertical input as well as a lateral and vertical output (see Figure 9). Overland flow

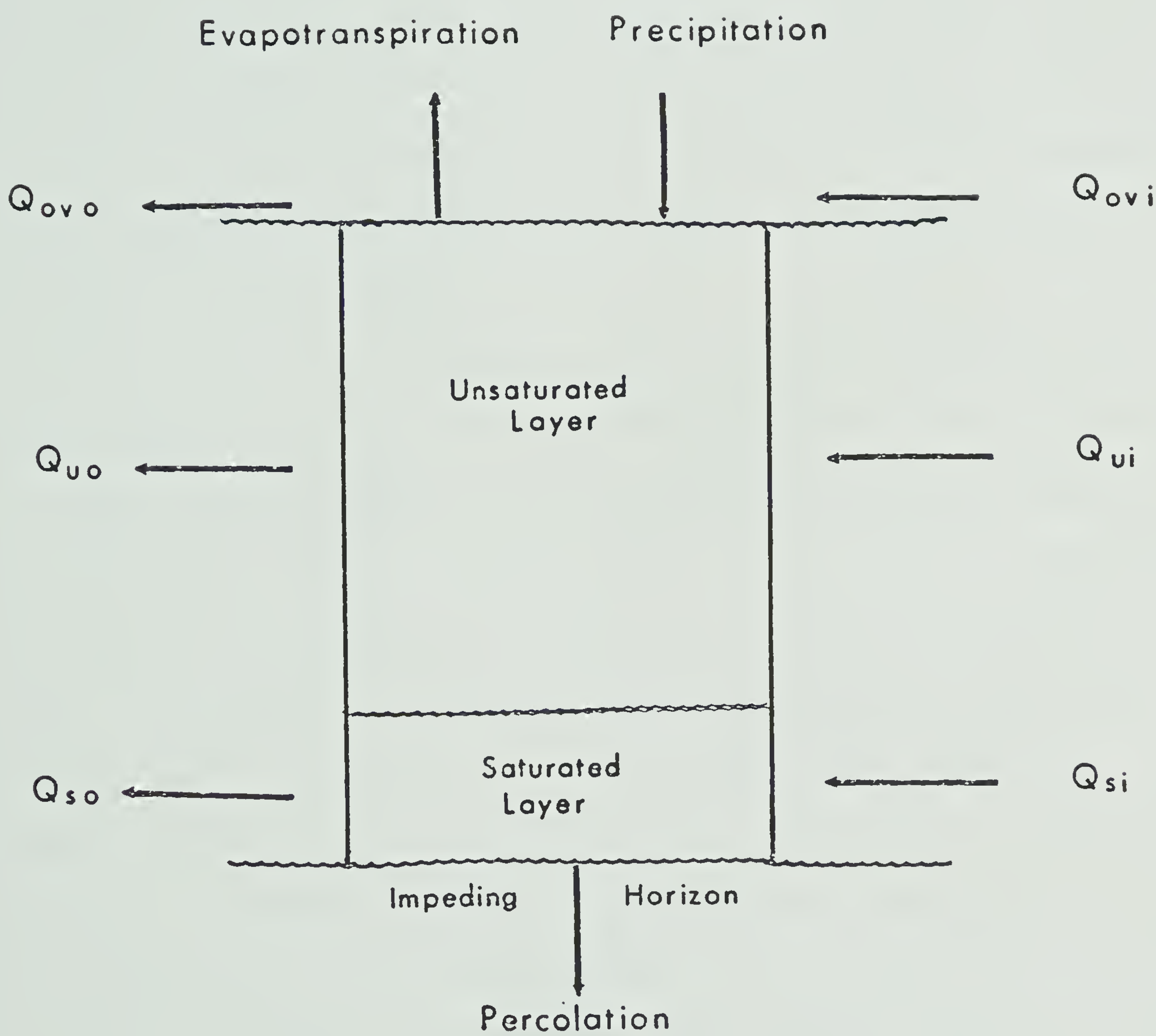


Figure 9 SLUICES model inflow-outflow diagram

storage has precipitation as an input as well as possible overland flow from adjacent elements (Q_{ovi}) and is depleted by lateral outflow (Q_{ovo}). The unsaturated storage can be decreased by lateral unsaturated flow from adjacent elements (Q_{ui}) and by lateral saturated flow (Q_{si}) and increased by lateral flow (Q_{uo}) out of the element. Saturated storage is increased by lateral saturated flow (Q_{si}) and unsaturated flow (Q_{ui}) from adjacent elements and decreased by outflow of this type (Q_{so}). Outflow in the form of deep percolation out of the saturated flow storage is also considered, representing the advance of a wetting front into the less permeable layer.

The amount of water in each storage is defined by a depth. Depth of water in overland flow storage denotes the depth of water that is available for overland flow while the depth for the other two storages denotes depth of soil. Let Y_u represent the depth of unsaturated soil and Y_s the depth of saturated soil. These two depths characterize the soil profile since only a certain portion (porosity) of the soil can hold water. Since an infinite number of combinations of air and water can make up porosity, an infinite number of saturated and unsaturated depths can make up the total profile depth. If no portion of the profile in the grid element is saturated (the moisture content is less than field capacity), then Y_s equals zero and thus Y_u equals Y, the total profile depth. However, if the moisture content is brought above field capacity, a saturated layer will be formed, decreasing Y_u. When the profile is completely saturated, Y_s equals Y and Y_u equals zero. Any increased input to a saturated

profile creates overland flow, which is quantified by a depth of water.

6.2.3 Calculation of Flow Quantities

6.2.3.1 Interflow

Calculation of the lateral, saturated flow quantity from any given grid element (Q_{sc}) is based on Darcy's law:

$$Q_{sc} = \text{Conductivity Coefficient X Gradient X Depth X Width X Time Interval}$$

The width of the element is fixed when the choice of element size is made. The hydraulic gradient is considered to be the difference in elevation between the top of the saturated storages in adjacent elements divided by the distance between them. This gradient is thus largely affected by watershed topography. The elevation of the saturated layer is the sum of the base elevations for the elements in question and the saturated depths for these elements. A check must be made each time this flow calculation is undertaken to ensure that the quantity of flow capable of being transmitted does not exceed that which exists in saturated storage. The amount of water in storage, Q_{sp}, is:

$$Q_{sp} = \text{(Saturation - Field Capacity) X Depth X Area}$$

with the area equal to the width of the element squared since the elements are square. This equation recognizes that saturated flow will deplete the soil water storage only to the level of field capacity. If Q_{sc} exceeds Q_{sp} then the saturated layer is depleted, discharge during the time interval from the elements is set equal to the volume of water in storage and the moisture

content is set at field capacity. If Q_{sc} does not exceed Q_{sp}, then the discharge equals Q_{sc} and then the saturated layer is reduced by an amount corresponding to the quotient of Q_{sc} and the area of the element. Inherent in these discussions is the assumption of a specified time interval. This interval will be discussed in greater detail later.

6.2.3.2 Overland Flow

When saturation of the soil profile and the subsequent formation of an overland flow layer occurs, the depth of this layer is assumed to be uniform over the entire element. Calculation of the overland flow component is based on Manning's equation:

$$Q_{ovc} = \text{Depth}^{1.67} \text{ X Width X Slope}^{0.5} \text{ X Time Interval/Manning's n}$$

The slope is calculated by dividing the difference in surface elevation (sum of the base elevations and profile depths) between adjacent elements by the distance between elemental centers. The depth of overland flow would generally be very small relative to the base elevations under consideration.

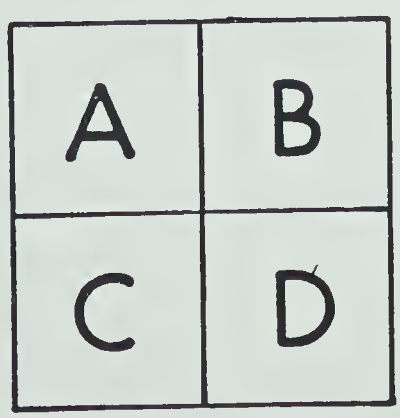
A check must be made to ensure that the quantity of flow capable of flowing out in the selected time interval according to Manning's equation (Q_{ovc}) does not exceed the amount in storage (Q_{ovp}). This latter storage is the product of the depth of flow and the area of the element. If Q_{ovc} exceeds Q_{ovp}, then the storage is depleted and the overland flow discharge during the time interval is set equal to the quantity of water in storage. If Q_{ovc} does not exceed Q_{ovp} , then the overland flow depth is

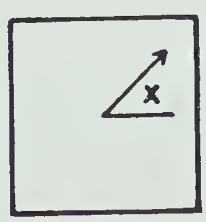
reduced by an amount corresponding to the quotient of Q_{ovc} and the area of the element and the discharge equals Q_{ovc}.

6.2.4 Flow Direction

Prior to the actual calculation of either overland flow or interflow, a subroutine of the model determines, for the element in question, which adjacent element to use for the calculation of hydraulic gradient for interflow and slope for overland flow. The choice of the adjacent element for slope and gradient calculations is based on an assumed slope direction of the element in question, based on an examination of a topographic map of the watershed.

Once an element size has been chosen, a grid system made up of this element size is overlain on a topographic map of the watershed in question. Elevations are assessed to each element found within the watershed. Assuming flow occurs perpendicular to the elevation contours, one decides, for each element, to which other element the slope is the steepest. This decision is translated to the model as an angle between 0° and 360°. Consider the four elements arranged as follows, with element A being the element in question.





For example, if the slope from element A were steepest to element D, then specification of an angle, X, for element A of between 292.5 and 337.5^{o} would result in the elevation of element D being used for the calculation of the slope for overland flow and the gradient for interflow for element A. If instead, the slope from element A was steepest to element B, then an angle between 337.5 - 360^{o} or 0 - 22.5^{o} would result in the elevation of element B being used in the flow calculations. These angles need only be specified as multiples of 45^{o}, since no significance is attached to the magnitude of the angle, other than in the general way described above.

Once the gradient and slope (if necessary) and consequently the outflow amounts are calculated, the outflow is routed to adjacent elements. The model routes water only to those adjacent elements that are on the same row and/or column of the grid as the element in question. In the above diagram outflow would be routed to elements B and C but not to D. Water could reach element D indirectly through elements B or C, or both, however. The proportioning of the calculated outflow that goes to either B or C is dependent upon the relative slope (A - B) versus (A - C). The exact proportion of the total flow to either element is calculated by comparing the pertinent slope (A - B) or (A - C)) as a ratio to the sum of the slopes (A - B) and (A - C). For example, if the slope (A - B) was 0.20 and slope (A - C) was 0.30, then 40% (.20/.50) of the calculated flow from element A would be sent to element B while the remaining 60% would go to element C.

If the angle described above for specifying the element to be used for gradient and/or slope calculations was 0 or 360°, element B would receive all of the outflow from element A whereas if the angle were 270°, element C would receive all of the outflow from element A. These two cases are the extremes with an infinite number of combinations lying in between, depending on the slopes (A - B) and (A - C).

This technique is employed for all the elements proceeding row by row, column by column, until the complete grid has been covered.

6.2.5 Model Parameters

The model involves the following parameters:

(1) Saturation: moisture content at which all the soil pores are full; corresponds to soil porosity,

(2) Field capacity: moisture content which separates drainage water from plant available water,

(3) Wilting point: moisture content at which plants can no longer extract water from the soil,

(4) Profile depth: depth of soil from the soil surface to the impeding layer,

(5) Drainage coefficient: determines the proportion of the saturated layer that, as a wetting front, passes into the less permeable layer, and is no longer considered,

(6) Conductivity coefficient: the rate at which water moves through the soil in a lateral direction, denotes responses at the outflow to an input at the inflow

point,

(7) Manning's n: roughness coefficient in the overland flow component calculation.

Estimates of the first four of the above parameters can be obtained from research studies. As with any other model, initial moisture conditions must either be known or assumed for each simulation run.

6.2.6 Effect of the Time Step

Basic to the operation of the model is the assumption of a time interval. This choice is based largely upon the frequency of meteorological data recordings which are usually made on an hourly or a daily basis. Considerations of evapotranspiration on a time interval of less than one day are generally inaccurate. However, consideration of interflow or overland flow on such a large time interval would be equally inaccurate, particularly for overland flow, which is characterized by large flow velocities. Water would move great distances in that large a time interval.

To determine the most appropriate time step (which is the number of equal increments that the time interval can be broken up into), computer programs were written to examine the effect on calculated outflow for both interflow and overland flow of changes in parameters affecting these flows, on a hillslope 2000 m long and 400 m wide. In each case a combination of factors was examined that would limit the accuracy of the result. A data input time interval of one hour was assumed.

6.2.6.1 Overland Flow

The factors which could conceivably affect the choice of a time step for overland flow are the depth of flow, width of the element, slope and Manning's n. The last two factors are characteristic of the particular element of the basin being considered and thus the ratio of slope$^{0.5}$/Manning's n (hereafter called the slope/n ratio) was considered as one factor. A maximum and minimum value for each were considered, thus bounding the conceivable range over which the ratio could vary. Slope was allowed to vary between 0.4 and 0.01 while n was varied between 0.03 and 0.10, giving the slope/n ratio a range of values from a maximum of 20 to a minimum of 1. Flow depths of 0.001, 0.005, 0.010, 0.050, 0.10 and 0.15 m were examined for element widths of 50, 100, 200, and 400 m. Another factor, the number of time steps was introduced. Values of 1, 2, 6, 12, 30, 60, and 120 were used. For example, use of 60 time steps would mean that the time scale of 1 hour would be broken up into 60 equal increments, implying that the calculation of overland flow is actually made on a one minute basis. One must recall that the potential amount of overland flow capable of occurring cannot exceed the amount of water in overland storage (depth of flow times element area). Thus any, or all, factors tending to maximize the potential overland flow would likely affect the choice of the time step. One might expect the choice of the time step to be most critical for the combination of factors that are maximum, for example, a depth of 0.15 m, element width of 400 m, and a slope/n ratio of 20. Because one can control the

size of the grid element, the choice of time step will be a function of element size. The other two factors of depth and slope/n ratio will depend upon the location of the element under consideration within the basin as well as the storm intensity and duration. These factors cannot be controlled. The choice of an appropriate time step will be based on the outflow value becoming a constant, thus implying that time step is no longer a factor in the calculations (which it should not be).

The results of this analysis on the effect of time step show that use of a single time step can cause errors in the calculation of outflow by as much as 29%. This error is a maximum and would be experienced in elements characterized by low slope, large n, large elemental width and large depths of flow. Time steps of 60 - 90 appear adequate under these conditions. Use of these time steps would correspond to calculating overland flow every 1 minute - 40 seconds for a time scale of 1 hour. The results also show that these requirements can be relaxed if the elemental size is reduced. For an elemental size of 100 m X 100 m or less, little benefit is gained by using time steps exceeding 60.

If greater flexibility in programming is desired, the time step to be used could also be made a function of the depth of flow. Depths of flow greater than 0.10 m require a time step of 60 or more but depths of less than 0.05 m require time steps of less than 30. This variation of the time step could prove beneficial since use of a larger number of time steps causes the computer simulation to be more expensive because more iterative

calculations are needed.

6.2.6.2 Interflow

The effect of the choice of time step was also examined for interflow, in a fashion identical to that for overland flow. The factors affecting the magnitude of interflow are conductivity coefficient, saturated depth, hydraulic gradient and element width. The difference between saturated moisture content and field capacity must also be considered since this difference will determine the amount of water actually available for interflow, since water held at moisture contents less than field capacity is not available for interflow. Saturated flow depths considered varied from 1.0 m to 0.01 m, with intermediate values of 0.50 and 0.10 m. Conductivity coefficient values examined were 1, 50, 100, and 500 m/hr as were hydraulic gradients of 0.10 and 0.50. Element widths of 50, 100, 200, and 400 m and saturation-field capacity differences of 5, 10 and 20% were utilized. Time steps of 1, 6, 12, 30, 60, 90 and 120 were considered.

Element width, as suggested in the discussion of overland flow, is a parameter which can be judiciously chosen by the modeller. Similar to the results for the overland flow analysis, results for interflow showed that the time step becomes most critical for the larger element sizes, with the error associated with a time step of 1, as opposed to a time step of 60, being a maximum value of 29%, with flow values for time steps of 60, 90 and 120 showing very little difference. Results using an element size of 50 m X 50 m were very similar, no matter what the time step.

Other results indicate that for interflow, a greater number of factors must be considered but that for element sizes greater than 100 m, hydraulic gradients steeper that 0.20, conductivity coefficients greater than 50 m/h, saturation-field capacity differences of 0.10 or larger and saturated depths greater than 0.5 m, use of a time step of 60 should provide adequate results.

6.2.7 Inherent Assumptions Made in Model Development

Certain assumptions were made in model development that were meant to simplify the model. The two major ones relate to infiltration and unsaturated flow.

6.2.7.1 Infiltration

Since the model will be applied to forested watersheds that have infiltration rates which exceed even the most intense storms, an assumption will be made that all the rainfall that reaches the soil surface infiltrates it. Beke (1969), in his study of the soils of three watersheds in Alberta, found that the mean minimum infiltration rates of the soils were generally higher than the maximum rainfall intensities reported, suggesting that the assumption should not be a limiting one. Many research papers relative to forest hydrology mention the absence of overland flow and by necessity, this implies that infiltration capacity is not a limiting factor.

6.2.7.2 Unsaturated Flow

The conductivity coefficient can be expected to be a strong

function of moisture content, decreasing dramatically as moisture content decreases, usually over several orders of magnitude. The result is the occurrence of very low flow velocities and consequently low flow quantities as moisture content decreases from saturation. For this reason, Anderson and Burt (1977) suggested that unsaturated flow is unlikely to be of any great consequence relative to the other flow processes being considered. Based on this belief, unsaturated flow was assumed to be negligible, with saturated interflow and overland flow being the two flow processes modelled.

6.2.7.3 Moisture Redistribution

Once water is added to the soil profile, it is assumed to be redistributed over the whole soil profile instantaneously. In actuality, the influx of water would proceed as a wetting front down the soil profile. This redistribution would take a finite amount of time. However, considering the rather shallow profiles being considered (less than 1 m in depth), the amount of time involved here, relative to the time required to travel the much greater lateral distances, is of minor importance.

The soil moisture redistribution process in nature is continuous. Its rate decreases constantly and equilibrium is reached only after very long periods of time. Thus one has difficulty in determining when redistribution has ceased. The presumed moisture content at which internal drainage ceases (field capacity) is thus a rather subjective and fictitious term. However, the field capacity concept is most tenable for

coarse-textured soils since hydraulic conductivity drops very steeply with decreases in moisture content and flow becomes slow relatively quickly. Despite its many shortcomings, the field capacity concept is still a useful and practical criterion for the upper limit of soil water content before drainage occurs, suggesting that water held below a certain moisture content is not available for redistribution.

6.2.8 Logistics of the Model

To facilitate the understanding of the logistics of the model, a generalized flow chart of the model is presented in Appendix II. The model has two general sections: one to calculate storage potential for soil moisture, depth of the saturated layer and overland flow, and the second to calculate the actual outflow and to decrement the relevant storages. Input to the model consists of precipitation and potential evapotranspiration. Watershed parameters which are initialized for each element are the saturated depth and/or moisture content. Characteristics of the elements of the watershed that are input into the model are the elevation of each element and the angle for each element (which was used to calculate the element(s) to receive the outflow). Model parameters to be given initial values include the six soil parameters of saturation, field capacity, wilting point, conductivity coefficient, drainage coefficient, and depth of profile. These parameters may or may not differ from element to element. The grid elements are taken to be horizontal and square. Output information includes soil moisture percentage,

saturated depth, actual evapotranspiration, drainage and the pertinent outflows from each of the elements.

CHAPTER 7

RESPONSE OF THE MODEL

7.1 Sensitivity Analysis of Model Parameters

Sensitivity of any model is a very important characteristic of that model. One can imagine that a model could be developed that would be acceptable since it met all of the other criteria previously outlined, but yet would not be sensitive to changes in parameters that correspond to land use manipulations. Such a model could not be used for predictive purposes related to land use manipulations.

Of course, the term sensitivity is subjective and thus it is used in a general sense. Implication of the term is that the model should react in a certain way, and within a certain range, to changes which have been shown to, or are suspected of showing, a certain effect. For example, for the case of clearcutting in forestry, it has been shown that such an action results in increased peak discharge and total volume of flow. Any model used for this purpose which would not show these trends would have been ineffective. Accuracy must also be considered, since if the model is sensitive to a given change, but in an amount less than the accuracy of the corresponding physical measurement of that parameter, use of that model is without justification.

Therefore, it is of prime importance that the effect upon

output due to changes in the parameters of the proposed model be investigated. Such an investigation helps to understand how the model operates and what its most important parameters are. In this way a fair assessment of the sensitivity and thus the suitability of the model can be made. One might expect that the three storage parameters characterizing the soil moisture reservoir should play an important role in determining the proportion of rainfall input that will be stored as soil moisture and the proportion that will occur as outflow. Other parameters will also play an important role in determining this proportion. One of these is the conductivity coefficient of the soil profile. A high coefficient would enable a large portion of the saturated zone to be depleted during a given time period and would cause this zone to be depleted much more quickly than one characterized by a low coefficient which would tend to keep more moisture in storage. Another important parameter in this regard is the drainage coefficient which allows a downward flow component of moisture out of the saturated zone. A large coefficient would tend to cause the saturated zone to be depleted more quickly than would a low one. This undoubtedly would have an effect on the shape of the recession limb of the hydrograph.

Another parameter to be considered is the profile depth. A deeper soil profile would mean that, because the moisture reservoir is larger, precipitation input would form a lesser proportion of the overall moisture in storage as opposed to that for a shallower profile, and thus, for the same given amount of precipitation in each case, outflow would begin sooner in the

case of a shallower profile, other things remaining equal. Also profile saturation would occur sooner, allowing the outflow to become at least partly overland flow sooner. This could have dramatic effects upon the hydrograph peak and shape.

A general discussion about the nature of the model parameters has been presented but a sensitivity study of the given parameters must be undertaken to be able to quantitatively assess the effects each can have on the output. This study generally involves varying one parameter while all others are kept constant. Because all of the aforementioned parameters have physical meaning, the range over which they should vary is limited by the range of the values normally experienced. Such a sensitivity study was conducted using the range of values deemed appropriate for the individual parameter. A 'watershed' slope was considered for the sensitivity study and consisted of a column of 5 grid elements, each of size 400 m X 400 m. At this stage only a slope was modelled to ensure that idiosyncrasies of element to element water routing were not affecting the results. By choosing a slope, the water could be routed in a straight line. The particular grid size was chosen because it corresponded to the size of commercial cut-blocks used in forestry operations in Alberta. The watershed slope was assigned a uniform, straight slope of 20% and a trial period of 31 days was used with outflow calculated on a daily basis. The meteorological conditions for the simulation trials were as follows: evapotranspiration for a period of seven days, two days of consecutive rainfall inputs of 25 mm each, on the eighth and

ninth days, and then evapotranspiration for the remaining 22 days. Potential evapotranspiration values were taken from evaporation pan data from Marmot Creek Basin, Alberta. In the sensitivity study evapotranspiration was modified by moisture content only. Evapotranspiration was reduced linearly from potential evapotranspiration (obtained from pan evaporation data) to zero as moisture content dropped from field capacity to wilting point. For moisture contents at field capacity evapotranspiration proceeded at the potential rate. The first seven days were used to check on the budgeting of the model and to represent antecedent conditions since the soil moisture was initialized at field capacity. The last 24 days were used to evaluate the shape of the outflow hydrograph. Each of the parameters was subjected to a sensitivity study to evaluate the consequences on the outflow hydrograph of changing their values.

The parameter values used in the sensitivity study were as follows:

(1) Saturation moisture content: 0.60, 0.55, 0.50,

(2) Wilting point moisture content: 0.20, 0.15, 0.10,

(3) Drainage coefficient: 0.00, 0.05, 0.10,

(4) Conductivity coefficient: 4, 22, 40 m/h, and

(5) Profile depth: 1.0, 5.0, and 9.0 m.

Each of the parameter values was kept within a range that could normally be expected within the field.

Two primary effects on the outflow hydrograph were examined: the peak value and the value after 31 days which was used to characterize the recession limb of the hydrograph.

Values were compared to show the trend established when the parameters were changed and just how sensitive the output is to changes in these parameters.

The results can be summarized as follows:

(1) A decrease in the saturation moisture content causes a significant increase in peak outflow and the outflow after 31 days.

(2) A decrease in the wilting point moisture content causes an extremely modest decrease in both the peak outflow and outflow after 31 days (both effects probably are not significant).

(3) A decrease in the drainage coefficient has no effect on the initial discharge but causes a significant decrease in the discharge after 31 days.

(4) A decrease in the conductivity coefficient causes a significant decrease in both the peak outflow and the outflow after 31 days.

(5) An increase in profile depth causes a very modest decrease in both the peak discharge and discharge after 31 days.

Assessment of the effects upon discharge of varying field capacity cannot be carried out in a manner similar to that used for the preceding 5 parameters because the initial moisture content had been set equal to field capacity. For the analysis of each of the preceding 5 parameters, the initial moisture content was set at field capacity. Initial moisture content cannot now be set at field capacity because, since field capacity itself is

varying, then initial moisture content would also vary, complicating the analysis due to the variation of two parameters instead of one. A choice exists as to whether initial moisture content should be set above or below field capacity. It should be recalled that a moisture content above field capacity is, according to the model, equivalent to a portion of the soil profile at field capacity and a portion saturated, with the total amount of water being conserved. Therefore, if field capacity were to be varied from 0.30 to 0.40 with an initial moisture content of 0.40, the profile having a field capacity of 0.30 would begin with a saturated layer whereas the profile with a field capacity of 0.40 would begin right at field capacity with no saturated layer. Rather than introduce this complication, a choice of a low initial moisture content seems more appropriate. If an initial moisture content of 0.30 were used instead for a profile whose field capacity was being varied from 0.30 to 0.40, the profile with a value of 0.30 would begin at field capacity but the one with a field capacity of 0.40 would start at somewhat less than field capacity, but in both cases each would begin with exactly the same amount of water with neither experiencing a saturated layer due to initial input of water. Thus the initial moisture content was set at 0.30 for the sensitivity trials on field capacity.

Unfortunately, however, if this value of initial moisture content is used along with the meteorological data used for the analysis of the preceding five parameters for varying field

capacity of 0.30, 0.35 and 0.40, outflow would not be experienced for profiles having field capacities of 0.35 and 0.40. Outflow does not occur because all of the precipitation experienced (2 storms of 25 mm each) would be used to satisfy soil moisture. It should be noted that four storms of 25 mm each would be required just to bring the soil profile with a field capacity of 0.40 up to field capacity from the initial moisture content of 0.30, for a profile depth of 1.00 m. Unfortunately this would not give any qualitative information about the runoff hydrographs for the upper two values of field capacity. Therefore, the meteorological data were rearranged to introduce six storms of 25 mm each, but using essentially the same evapotranspiration data as before. Results showed that the profile with the lowest field capacity experienced the highest discharge rate throughout the thirty-one day period. Also discharge occurred soonest in the profile with the lowest field capacity, because not as much precipitation was being used to restore soil moisture defecits. This suggests that the peak discharge would occur sooner for profiles with low field capacities compared to profiles with higher field capacities. However, the exact nature of the response of a particular soil profile depends upon the value of the saturation coefficient as well.

The sensitivity study was subsequently repeated on a 10 element watershed divided into a grid that consisted of 4 rows and 3 columns with the upper 2 corner elements excluded from the watershed. Each element had dimensions of 400 m X 400 m.

Elevations for the elements were taken from a topographic map of Marmot Creek Basin in the Rocky Mountains in an attempt to utilize elevations, and thus slopes, that were characteristic of a mountain watershed. These slopes varied from 9% to 44%. The purpose of this exercise was to investigate whether or not routing of the outflow through a non-straight path might have some hidden effect on the results. Such was not the case and the effects of the sensitivity study reported for the 5 element slope are practically identical to those experienced for the 10 element watershed.

Another important factor besides discharge rate that should be considered is soil moisture, mainly because of its availability or nonavailability for plant use. It also is a measurement often made in the field and thus can also be used as a check on the accuracy of most models of the hydrologic cycle. Since the value of soil moisture content would be greatly dependent upon the location of the element in question within the watershed, each of the elements of the watershed being modelled should be considered. Results obtained for a short term discharge period for the 10 element watershed showed the actual variation in moisture content between elements was rather small. This was probably due to the following reasons: (1) the short time period being considered (31 days), and (2) consideration of similar evapotranspiration for all elements that is dependent only upon soil moisture. However, the differences in the saturated depths between elements were marked. Those elements with the lowest gradients experienced the highest saturated depths while elements

which were characterized by the highest elevations and the highest gradients experienced the lowest soil moisture content and the lowest saturated depth.

Also of interest are changes in soil moisture due to changes in parameters. One case is chosen to represent the trends established in moisture content changes due to parameter variations. This case is the variation of saturation percentage from 0.60 to 0.50. Recall that this change in saturation percentage resulted in a significant increase in peak discharge. Differences in soil moisture were minimal but the saturated depths for a saturation percentage of 0.60 were consistently higher than those for a saturation percentage of 0.50. This suggests that the higher peak discharges occurred at the expense of the saturated layer. This seems logical since the saturated layer is the source of the discharge. Also for a field capacity of 0.40, a saturated layer in a soil profile with a saturation percentage of 0.60 would release twice as much drainage water as would an equivalent saturated layer in a soil profile with a saturated percentage of 0.50. This fact accounts for the greater amount of moisture in storage in the saturated layer under a saturation percentage of 0.60 suggesting that the recession of the discharge hydrograph would be much more gradual under these conditions.

7.2 Effect of Varying Element Size

Another important consideration which must be made in the use of the model is the choice of element size. Obviously, for a

given slope or watershed, an infinite number of choices of element size is available. To investigate the effects of element size on hydrograph shape, modelling of a slope 2000 m long and 400 m wide was undertaken, with the same meteorological conditions as used in the sensitivity study. A uniform slope of 20% was chosen. Considering elements lengths (distance in the direction of flow) of 400, 200, 100 and 50 m and widths (distance perpendicular to the flow) of the same magnitude, a total of 16 possible combinations of the above lengths and widths are possible. Each of these combinations was modelled. The results showed that the width of the element had no effect on the outflow. This means that the outflow of a slope of 400 m in width could also be obtained by the sum of the outflow from two 200 m wide slopes, four 100 m slopes or eight 50 m slopes. However, the length of the element had a most noticeable effect. The differences among the hydrographs for the four slope lengths are evident from Figure 10, with the greatest effect being shown on the recession limb of the hydrograph. A larger element length causes a delayed recession compared to that of a smaller length. However, the principle of diminishing returns sets in and the hydrograph obtained using an element length of 100 m is not significantly different from that using lengths of 50 m. One should keep in mind that the choice of element length of 50 m requires twice as many elements as does the choice of a 100 m length, to cover the same area.

In the modelling of a watershed, water would routinely be routed to adjacent elements, both along the same row and the

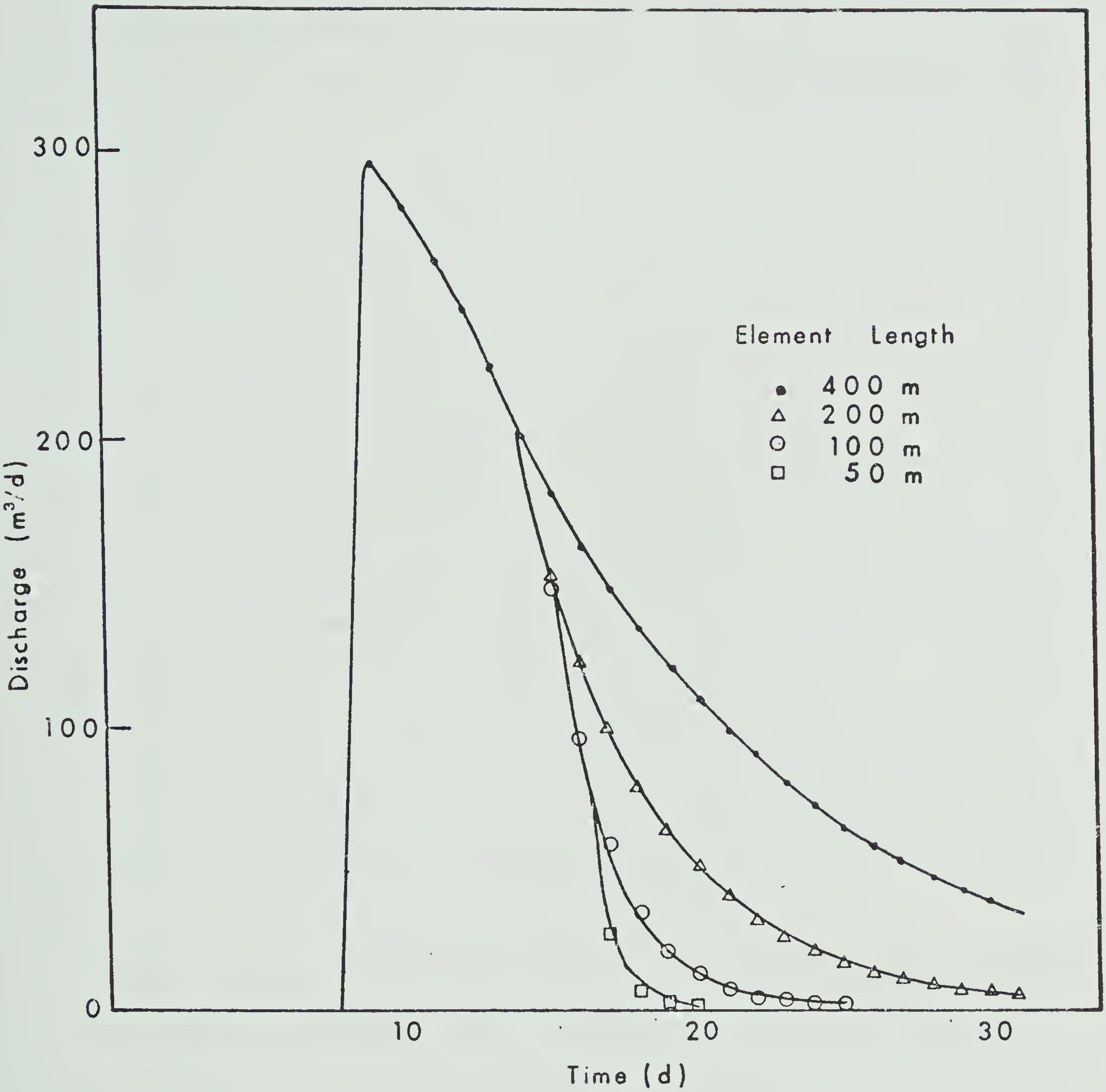


Figure 10 Sensitivity test on element size

same column as the element in question. Therefore, the definitions of length and width as given above would not hold true. However, the conclusion that element size should be made as small as is feasibly possible would still apply. Use of square elements would ensure that the effects of both length and width are accounted for.

The result that element size can affect the shape of the hydrograph should come as no surprise. The principle is similar to the evaluation of an integral using, say the trapezoidal rule, and finding that as the number of subdivisions is increased, a better approximation of the value of the integral is achieved. This effect also manifests itself in the model in the choice of the time interval. It has been shown that a more accurate result can be achieved by increasing the number of the time intervals but that the principle of diminishing returns also sets in giving little increased advantage to further increasing the number of time intervals. Again practicality in the modelling will help decide the time interval to be used much as it will the element size.

The results of changing element size just discussed suggest an important conclusion: the structure of the model places a significant limitation on the choice of element size. The results suggest that a modeller using the model should use an element size as small as feasibly possible, keeping in mind that an increase in the number of elements in the grid system will result in an increase in the computational burden on the model. The decision on element size cannot, therefore be based entirely on

physical characteristics of the area being modelled. For example, a large homogeneous area should still be described by a grid system using relatively small elements. However, in view of the fact that the model was developed for mountainous watersheds, large homogeneous areas are undoubtedly an exception rather than the rule. Thus, use of a small element size would probably be advisable from a topographic point of view. In addition, if one is considering numerous watershed characteristics, for example, topography, soils, vegetation, etc., homogeneous units would be of rather small area anyway. In addition, the handling of distributed inputs to the model, such as precipitation or moisture content, would be facilitated if the grid elements were smaller.

However, this is not to suggest that, since modern computers can handle extra computational burden with relative ease and only slight added cost, a modeller is not limited by how small the elements can be made. One must remember that each individual element must have an elevation, obtained from a topographic map, specified for it. As a result, the scale of the topographic map must be considered. It is pointless trying to obtain elevations for individual elements from a topographic map whose scale is too small, relative to element size, to allow the specification of reasonably accurate elevations. Thus the topographic map used for specifying elevations must be the ultimate source of limitation to element size. Therefore, element size should be made as small as physically possible, based on the largest scale of topographic map available for the watershed being studied.

C H A P T E R 8

VALIDATION OF THE SLUICES MODEL

A sensitivity study is important in the assessment of how a model responds to changes in its variables. Validation, another important step in the assessment of any model, is an attempt to see if the model can provide known outputs for given known inputs. For this reason, model validation can said to be an assessment of the truth of the model.

8.1 Comparative Validation

One technique of model validation is the comparison of results from the model under investigation to one that has already been recognized.

8.1.1 The Freeze Model

A suitable model for comparison purposes is that developed by R. A. Freeze which coupled unsaturated flow with saturated flow utilizing the generalized partial differential equations for each respective flow system and a numerical finite difference method for problem solution. Functional relationships between hydraulic conductivity, specific moisture capacity, moisture content and pressure head were used as input using tables of values to represent the wetting and drying cycles included in hysteresis. Needless to say, the model required a great abundance of data. (For greater detail, see Freeze, 1972b).

Freeze's application of his model to an investigation of the role of subsurface flow in generating runoff in upstream areas is of interest here. He attempted, using the model to answer the questions: "Is subsurface stormflow important quantitatively?" "Is it simply a controlling mechanism on wetland areas?" or "Is it unimportant in all contexts?"

In his model simulations, Freeze examined the effects of variations in certain physical parameters on runoff generation on a rectangular hillslope 120 m long X 33.5 m wide. The parameters (suitable for comparison in this discussion) that were investigated and the range over which they were allowed to vary were as follows:

(1) Saturated hydraulic conductivity (0.44 - 0.00044 cm/sec),

(2) Soil thickness (20 - 200 cm),

(3) Hillside slope (Gross slope - 7.5 and 15%), and

(4) Shape of the slope (Convex and concave).

The conclusions drawn by Freeze using his model are noteworthy and several are examined below:

(1) A necessary condition for the dominance of the subsurface flow mechanism is a convex hillslope.

(2) The saturated hydraulic conductivity of the soil exerts a greater influence on the runoff generating system than does the soil-slope configuration.

(3) For any soil-slope configuration there is a threshold saturated hydraulic conductivity value below which subsurface stormflow is not a feasible mechanism of

runoff generation.

(4) Changes in soil thickness do not affect the hydrograph significantly.

(5) The steeper sloped hillslopes produce higher downslope hydraulic gradients in the soil but this tendency towards larger outflow is offset by the need to recharge the lower moisture contents that exist in the steeper slopes under static initial conditions.

(6) Considerable evidence to support the claim that horizons of shallow surface soil of high hydraulic conductivity are a common occurrence in both forested and agricultural watersheds exists, but such soil conditions do not guarantee the eventuality of subsurface stormflow. (Freeze, 1972b).

8.1.2 The SLUICES Model

In an effort to compare results with Freeze's model, slopes of various shapes and degrees were examined with the SLUICES model. The slope shapes investigated were concave, straight and convex. Degrees of slope investigated were 10, 20 and 40% (all gross slopes). The slope consisted of 5 elements in a row, each 400 m X 400 m in size. All other parameters were set to those corresponding to initial conditions in the sensitivity study discussed earlier. These parameters were the same for each element. The same meteorological conditions and time span were used for this study as for the sensitivity study.

8.1.3 Comparison of Results

The conclusions which can be drawn through the use of the SLUICES model for this theoretical hillslope compare very favorably with those drawn by Freeze using his model. These similarities can be summarized as:

(1) Both models demonstrate that the hydraulic conductivity (conductivity coefficient) is the most important parameter affecting interflow and that low values of this parameter cause interflow to become insignificant.

(2) Both models show that soil thickness does not affect the hydrograph significantly.

(3) Both models show that interflow is most dominant on convex slopes and rather insignificant on concave slopes.

8.1.3.1 Hydraulic Conductivity

The conclusion that must be considered central to the discussion of the role of interflow is the one that saturated hydraulic conductivity is the parameter having the greatest influence on interflow. Both models support this conclusion. However, Freeze disregards the largest value of saturated hydraulic conductivity that he utilized (380 m/day) as being unrealistic for two reasons: (1) Use of this saturated hydraulic conductivity value causes almost 100% of the precipitation to be delivered to the channel and discounts this runoff event as

being unrealistic, and (2) saturated hydraulic conductivity values of this order of magnitude have yet to be supported by field evidence. His second justification for their exclusion is the one most subject to controversy. Freeze discusses laboratory measurements of saturated hydraulic conductivity in support of his view. The stand has already been taken in the development of the SLUICES model that the conductive properties of the soil will be characterized by a coefficient which includes not only the transmission properties of the soil matrix (usually measured in the laboratory) but also those properties which result from the presence of animal burrows and root channels and tend to increase the transmission rate dramatically. Forest hydrologists are cognizant of this fact even though they are willing to concede that it is difficult to measure a representative transmission rate in the field considering the likely dramatic spatial variations of such properties.

Hewlett (1974) considered Freeze's remarks about saturated hydraulic conductivity 'unsettled' since, as he points out, Freeze has assumed that the average and range of hydraulic conductivity values that would normally be measured in field soils would be adequate to characterize the saturated values of surface soils and stream bank materials. Hewlett suggests that channel banks would have hydraulic conductivity values that tend towards the highest limits set by Freeze. Data obtained by several researchers have shown that values for the conductivity as high as 4 m/h may not be unreasonable. For example, Harr (1977) reports saturated hydraulic conductivity values of

352 cm/h and 412 cm/h for the uppermost two soil layers in his study watershed in Oregon. Chamberlin (1972) working in a forested watershed in British Columbia found a hydraulic conductivity of 35 cm/h for the saturated basal zones, a hydraulic conductivity of 0.3 - 1.7 cm/h for the unsaturated B horizon but a total profile hydraulic conductivity of up to 200 cm/h. Beasley (1976) in an attempt to rationalize subsurface flows that peaked within one hour after rainfall began, surmises that flow velocities must have exceeded 33 m/h. He postulates that water travelled through macrochannels formed by decayed roots (a common occurrence in forested watersheds) and concludes that these macrochannels formed pathways for rapid movement of water. He suggests that any estimate of hydraulic conductivity would be in serious error unless some allowance is made for the effects of decayed root channels. Therefore, the apparently high values of saturated hydraulic conductivity used by Freeze should be considered as plausible, and not unconditionally rejected.

One must remember that the conductivity coefficient referred to above is actually an average value for the entire profile. Thus the extreme upper soil layers (and forest litter) might require even higher values of the coefficient to compensate for the lower ones characteristic of the more dense mineral soil below. The range of values experienced for the overall conductivity would obviously depend on the depth of profile considered.

8.1.3.2 Depth of Profile

One other conclusion which requires comment is that the profile depth was found not to be a significant parameter in the sensitivity study. Even at shallow depths, profile depth had very little effect upon outflow, as simulated by the SLUICES model. One should recall that in the sensitivity study, deliberate attempts were made to ensure that overland flow, resulting from profile saturation, did not occur in order to avoid comparison of overland flow and subsurface flow, or a combination of the two flows. Thus at no time was the profile saturated since all precipitation could be handled by storage in the soil profile.

Profile depth would however become significant in profile saturation. A shallow profile has less total water storage available than does a deeper profile, and as a result, a shallow profile would become saturated sooner (and thus overland flow would occur sooner) than would a deeper one, if enough precipitation occurs to saturate at least the shallow profile. Betson and Marius (1969) present results which support this view. They suggest that when an upper soil horizon is thin and the percolation rate of the next lower soil horizon is limited, saturation can readily occur, with saturated interflow occurring where the A horizon is the shallowest. The agricultural watershed (1.88 ha) in North Carolina that they studied had areas within it that had A horizons less then 7.5 cm in depth. Most watersheds would be characterized by variable soil profile depths throughout with the shallow soil profiles being located on

the ridges and deeper ones in the low-lying areas. The hydrologic implication of this general trend is that even though the low lying areas tend to receive water input both in the form of precipitation and subsurface flow, they are generally characterized by deeper soil profiles which are capable of storing more water before saturation of the entire profile occurs.

8.1.3.3 Slope Shape

Through the use of concave and convex slopes in his model, Freeze drew several conclusions in regards to interflow. These are:

(1) As saturated hydraulic conductivity decreases, subsurface stormflow decreases, and the rate of growth and the maximum size of the near channel saturated wetlands increases.

(2) Subsurface stormflow is quantitatively insignificant for all but the most permeable soil veneers. On concave slopes direct runoff from transient wetlands adjacent to channels dominates the hydrograph.

(3) Subsurface stormflow is also ineffective in raising the water table levels that control the size of the wetland areas. Surface saturation occurs because of vertical infiltration to the very shallow water tables rather than by downslope subsurface feeding.

Using the SLUICES model, three slope shapes, convex, straight, and concave, were investigated. Because convex and concave slopes, for a given gross slope, can take on different

degrees of slope, care was taken to ensure that the steepest portion of the slope did not exceed 40%, considered to be an average mountain slope. Results of the simulation for the slope shapes revealed that the convex slopes had the highest discharge while the concave had the lowest. Concave slopes were characterized by elements near the outflow point that had exceeding low slopes whereas convex slopes would be characterized by much higher values. Because of this, convex slopes had the highest outflow values and concave the lowest ones. However, concave slopes would also be characterized by water being retained for a longer period of time in the elements near the outflow point than would convex slopes. As a result, for any subsequent storm, less precipitation input would be required to recharge lower moisture contents on concave slopes and discharge may actually begin sooner on the concave slopes.

8.1.4 Summary

In view of the extremely divergent, independent approaches taken in the development of each of the two models and the rather dissimilar data requirements of each model, the qualitative support that the Freeze model gives the SLUICES model must be considered significant. Considering the attention given to the Freeze model and the success he reports in using it, one should be able to use the SLUICES model with some degree of confidence.

As has been mentioned previously, one important criterion in model selection must be simplicity. The model being developed,

while showing results very similar to those given by the Freeze model, is much more simply structured than the Freeze model. Freeze admits that because of the limitations imposed by computer storage and time and by the limited availability of the necessary data, the routine use of models like his would not come to pass in the foreseeable future, certainly not for field applications. Thus any model such as the one described herein, which can show results similar to those derived from more complex models, must be considered a welcome tool.

8.2 Watershed Simulation

One other technique for model validation is the direct application of the model to an instrumented watershed and the comparison between predicted and actual outputs for known specified inputs. The output most often utilized for comparison purposes is streamflow. Another output that can be used for validation purposes is soil moisture, if suitable data for its utilization are available.

In the validation procedure, it is important to apply the model to a watershed that is characterized by the flow processes of primary interest. For example, the SSARR model is best suited to larger watersheds with a well developed stream channel system. The SLUICES model was developed for use on basically first order watersheds whose flow processes are dominated by interflow. These watersheds are generally small, steep, forested ones. Since the model is based on the theory that it is predominantly lateral flow, caused by a flow impeding horizon,

that is the basis for interflow, validation of the model should be attempted on a watershed for which research has shown that interflow does indeed exist. Anomalies in results can be much more easily explained if complicating factors can be minimized. Such factors include variations in precipitation, particularly snow. Because of the wide variations in streamflow that can be created by variations in snow accumulation patterns and snowmelt rate, storms in the form of snow or rain-snow combination should be left for later analysis, after the model's treatment of the interflow process has been validated.

8.2.1 Test Watershed Description

Keeping the above criteria in mind, the Jamieson Creek watershed located in the Seymour River Basin near Vancouver, British Columbia, was selected for validation purposes (see Figures 11 and 12). This watershed, 3 km^2 in area, extends from 300 to 1300 m above sea level, has a southeasterly aspect and an average slope of 48%. Due to the proximity of the Pacific Ocean, the climate is typically maritime. The wettest months are generally October, November and December while the driest ones are June, July and August (Figure 13 depicts the precipitation distribution for Jamieson Creek). Rainfall accounts for about 90% of the 3910 mm of mean annual precipitation. Rainfall intensities rarely exceed 25 mm/h but storms are generally of a long duration, sometimes lasting several days. Snowfall, accounting for the remaining 10% of the precipitation, usually occurs from late November to mid-March (Cheng _et al._, 1977).

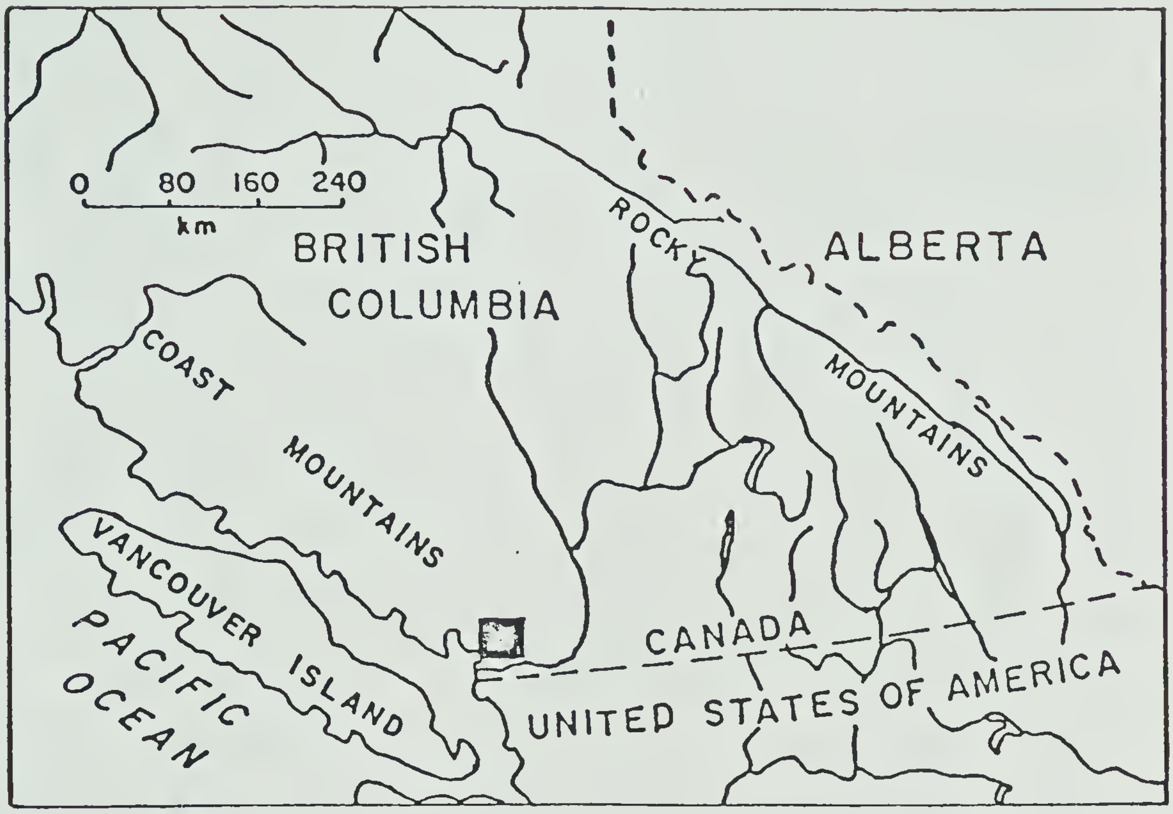


Figure 11 Location of the study watershed (Cheng, 1975)

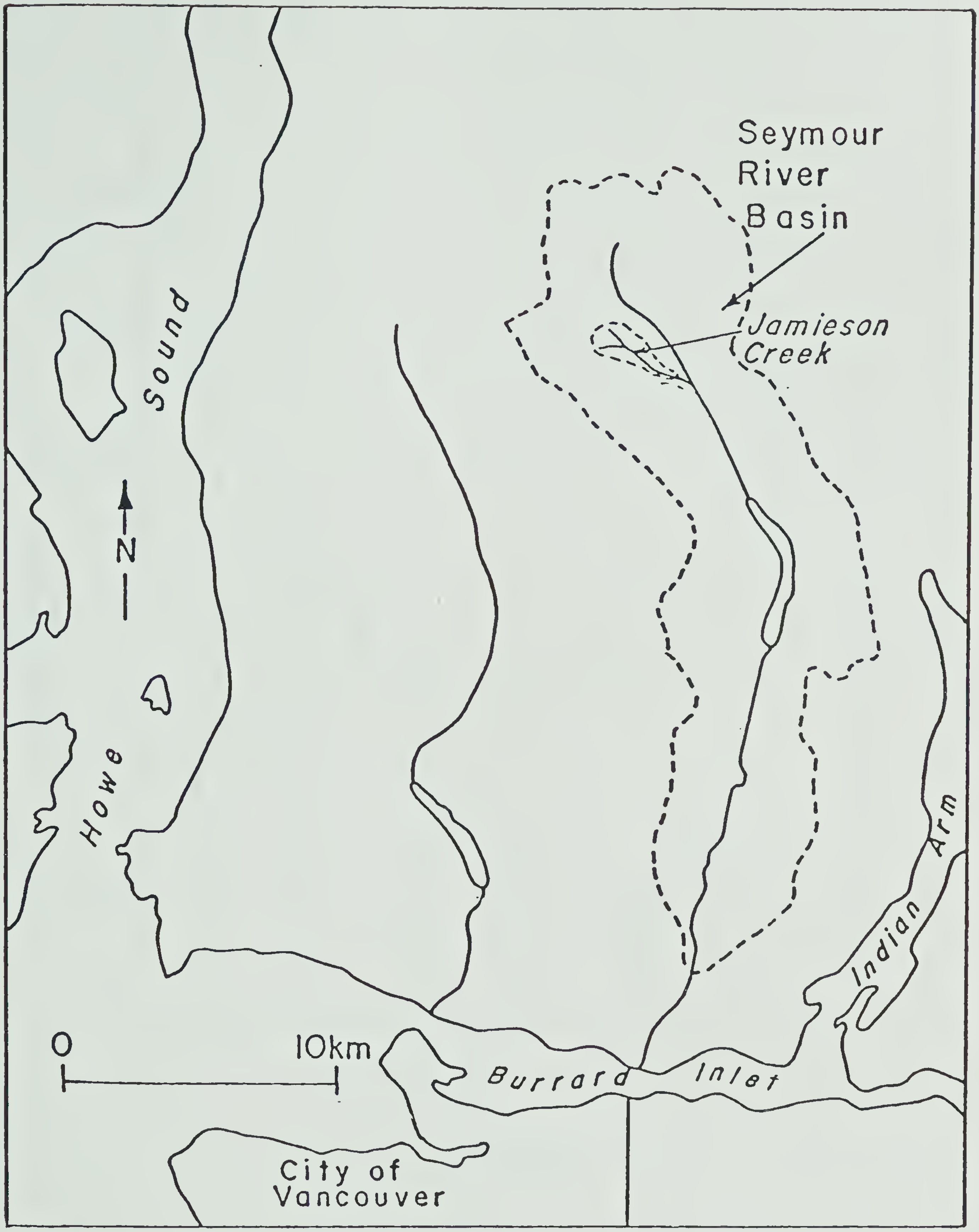


Figure 12 A map of Seymour River Basin, showing the location of Jamieson Creek (Cheng, 1975)

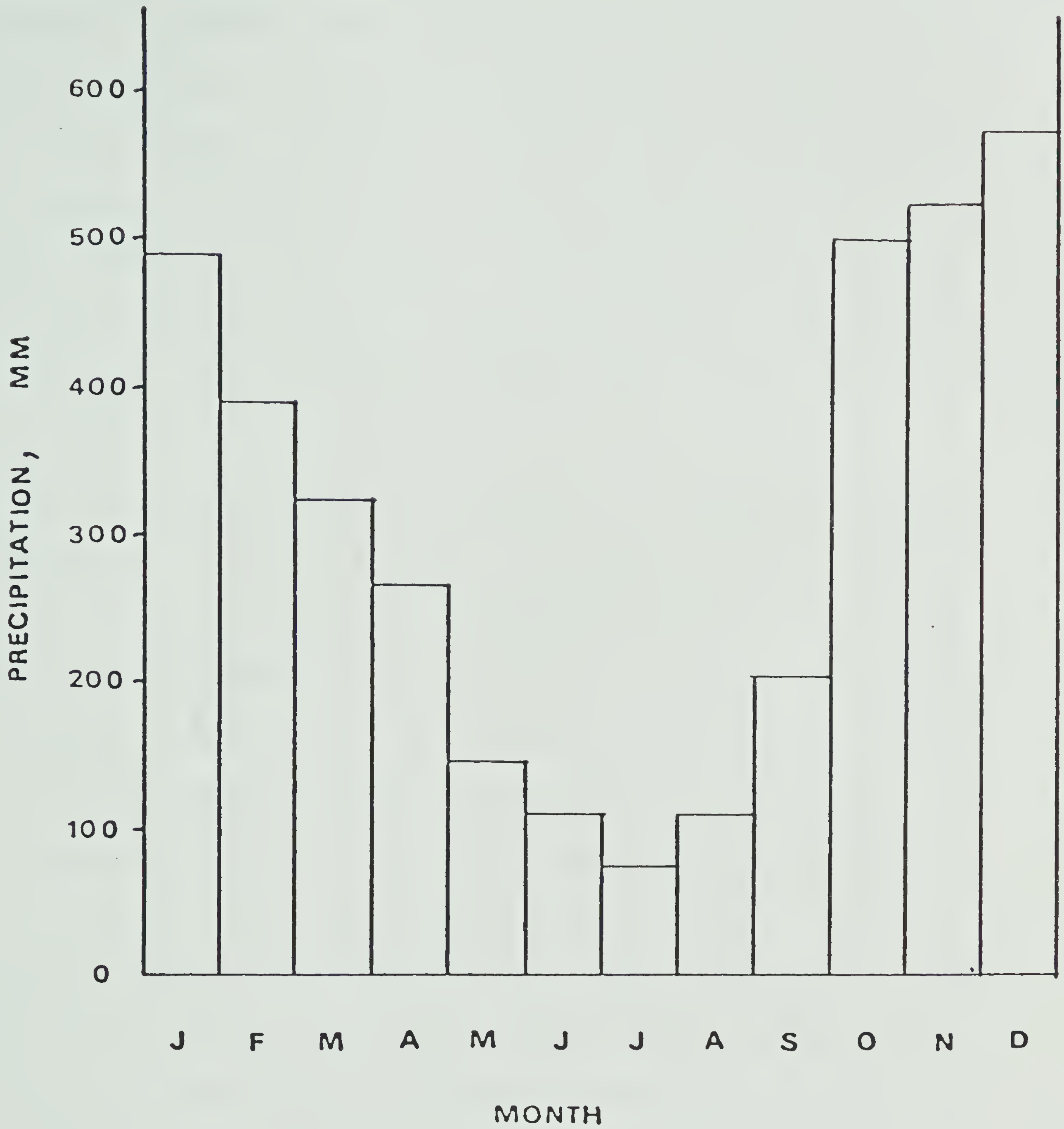


Figure 13 Monthly distribution of precipitation at Seymour Falls, Seymour River Basin 1941-1970 (Cheng, 1975)

Zeman (1973) described Jamieson Creek as a topographically well-defined watershed with relatively watertight bedrock and an undisturbed coniferous forest ecosystem. Jamieson Creek watershed is entirely covered by mature and over mature coniferous forest. Below an elevation of 900 m, in the Coastal Western Hemlock zone, the most productive trees are Douglas fir, western hemlock, western red cedar, sitka spruce and Pacific silver fir. Above 900 m the subalpine mountain hemlock zone begins with the main tree species being mountain hemlock and Pacific silver fir (Zeman, 1973).

A relatively impervious basal till is commonly encountered at depths of 0.5 to 1.2 m below the ground surface. Generally loamy sands with a high stone content predominate in the tills. The texture in the B horizon is finer as a result of weathering and gravelly sandy loams are generally common (Lewis, 1973). Depth of the surficial material to bedrock is largely controlled by topography. The till has mostly been removed by colluvial action from the steep upper valley slopes whereas it remains on the gentler slopes of the lower valley walls. Soils of the area are generally gravelly sandy loams derived from glacial till and are usually covered with a layer of thick porous forest floor (Plamondon _et al_., 1972). The experimental site examined by the authors was characterized by a 30 cm soil profile. The forest floor was 17 cm thick, with the L horizon being 1 cm thick, F being 7 cm thick and H being 9 cm thick. Bulk densities ranged from 0.09 g/cm^3 near the surface to 0.18 g/cm^3 for the H portion of the litter layer. An eluviated horizon (Ae), 3 to 5 cm thick,

underlays the forest floor, and contains many discontinuities caused by roots, stones and organic matter. The soil profile was developed on compacted glacial till.

Chamberlin's (1972) experimental site was located at an elevation of 700 m on a 30° slope. The soils in the area were generally Podsols, with thick organic layers (10 - 30 cm) in a well developed Ae horizon and a B horizon 0 - 1 m in depth.

Lewis and Lavkulich (1972) examined some shallow well-drained organic soil (Folisols) in the Vancouver area. They found that these soils were associated with steeply sloping bedrock and deeper mineral soils (Ferro-Humic Podsols). One site they examined on Mount Seymour was located at an elevation of 850 m on a long continuous steep (35%) slope with a south aspect. The 63.5+ cm profile was dominated by an organic LFH layer with a 1 cm loamy sand Ae horizon underlain by massive quartz diorite bedrock. The authors described several typical soil sequences that could be found in Jamieson Creek and reported the loss of elements such as calcium, magnesium and potassium to drainage water because of the frequent rapid flushing of the soil caused by the high permeability of the mineral soil, steep slopes and high annual precipitation in the watershed.

Willington (1971) considered the root zone to lie in the 0 - 60 cm depth range and the drainage zone in the 60 - 90 cm range. He found that the increasing bulk density below 60 cm of soil depth formed a barrier to the downward progression of roots and suggested a maximum rooting depth of 75 cm with the

maximum root density occurring between 30 and 60 cm.

Zeman (1973) distinguished the following three broad categories in the watershed, primarily on the basis of parent materials, depth over bedrock, profile development and drainage pattern from an hydrologic point of view:

(a) Shallow well-drained soils, usually in the upper part of the watershed, which developed on bedrock and are dominated by poorly to well decomposed organic matter horizons. The L, F, and H horizons are often in direct contact with the bedrock.

(b) Soils which developed on imperfectly drained sites and have formed in weathered basal till overlying unweathered basal till or bedrock under the influence of seepage water. These soils are periodically waterlogged and occupy lower slopes, midslope benches and drainage depressions. They have shallow L, F and H horizons underlain by a gleyed and stony Bhf horizon.

(c) Soils which developed in till and/or colluvium under well-drained conditions. These soils show good horizon development.

The low intensity rainfall, in combination with the permeable soils, suggest that overland flow as a flow process rarely occurs. Cheng et al. (1975) speculated that the overall permeability of the soils of Jamieson Creek should be greater than 200 mm/h. O'Loughlin (1972) estimated that the saturated hydraulic conductivity for a field soil in Jamieson Creek

containing channels might be as high as 670 - 2000 mm/h. Zeman (1973) stated that the very permeable soils, steep slopes and great annual precipitation result in frequent flushing of the soil. Feller (1975), in a study of five watersheds in the UBC Research Forest at Haney, B. C., described the response to precipitation as fairly rapid and hypothesized that stormflow arose mainly from the flow of water through macrochannels in the soil.

This watershed is well-suited for model validation purposes because of the belief that interflow plays a dominant role in discharge from the watershed (Cheng, 1975). All the above information about the watershed would seem to support that belief. As well the watershed supports a virgin stand of merchantable timber and, at the present time, is being logged for streamflow augmentation purposes by the Greater Vancouver Water District, under the guidance of the Faculty of Forestry at the University of British Columbia. Thus validation of the model at Jamieson Creek adds further relevance to it.

8.2.2 Model Preparation

8.2.2.1 Grid System

A map of the Jamieson Creek watershed (scale 1" = 400', contour interval 25') was overlain with a grid system having square elements of size 100 m X 100 m. The grid system required was a matrix 15 rows deep and 32 columns wide. The number of elements included within the watershed was 300, giving an

overall grid area of 3 km^2, which is the actual watershed area.

A topographic map of Jamieson Creek is shown in Figure 14. On it are marked the 100 m contour intervals as well as the 100 m X 100 m grid system utilized to divide the watershed into elements. From the topographic map of the watershed, five elevations, arranged as in the number five on a dice, were obtained for each element. For the 300 element grid system (100 m X 100 m elements), these five elevations were averaged to give the representative elevation for that element. Two other grid systems with different element sizes were then derived from the 100 m grid system. A grid system with element size 200 m X 200 m was obtained by combining four elements of the 100 m X 100 m size together. This grid system had 75 elements. Another grid system having element size 50 m X 50 m was obtained by dividing the 100 m X 100 m grid system into quarters giving 1200 elements to cover the entire watershed. For the 75 element grid system, the twenty elevations from four adjacent elements were averaged to obtain the elevation for the given element. For the 1200 element grid system, each of the four outside elevations, of the initial five, were specified as representing an element and the central elevation was ignored.

Flow direction angles were specified for each element for all grid systems. These were decided upon by consideration of a topographic map and the elevation contour lines on it. (Recall that the angle determined the element that was to be used in slope and gradient calculations.) All angles were specified to the nearest 45°. Figures 15a, 15b and 15c depict the three grid

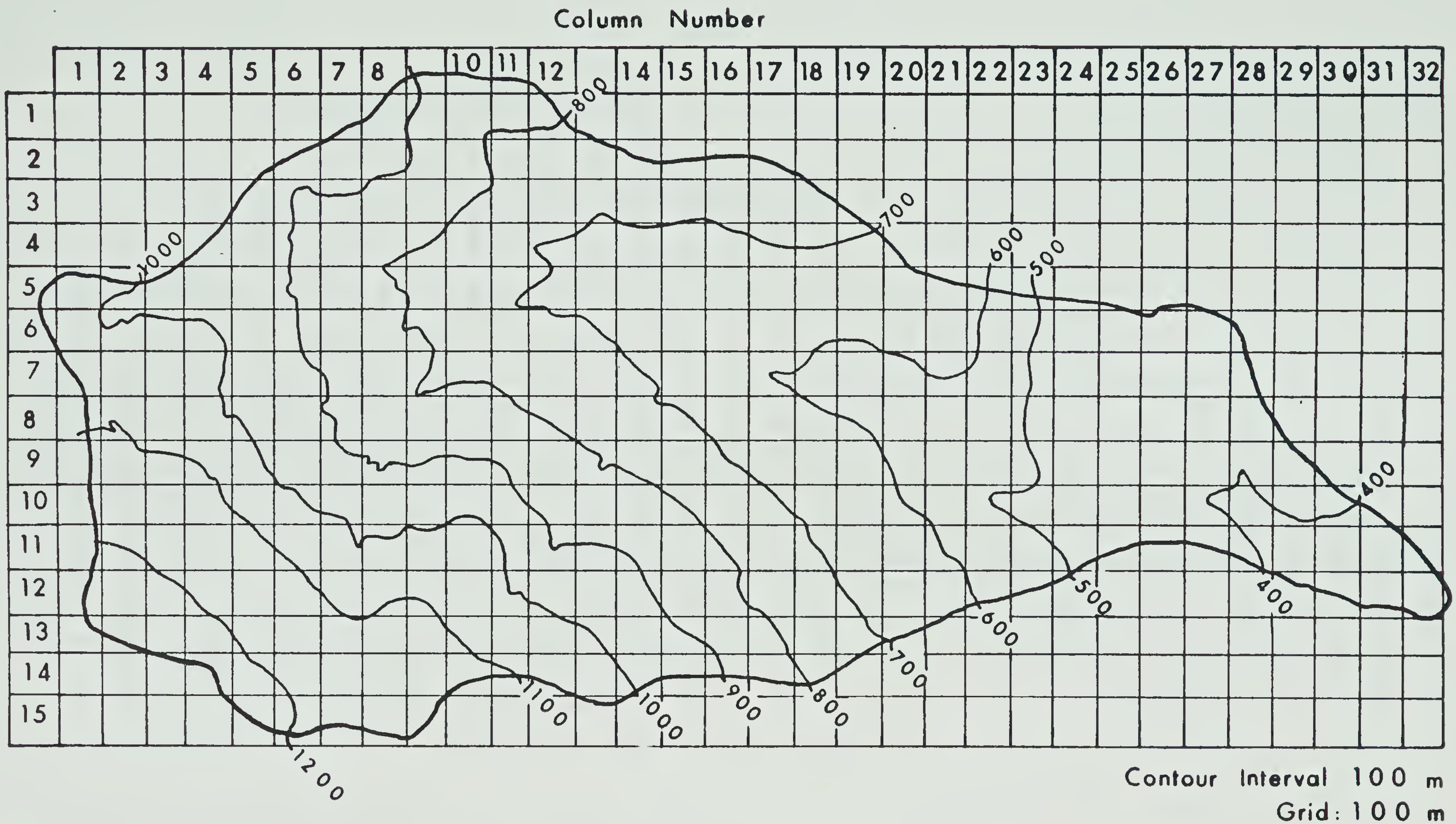


Figure 14 Topographic map and grid system for Jamieson Creek

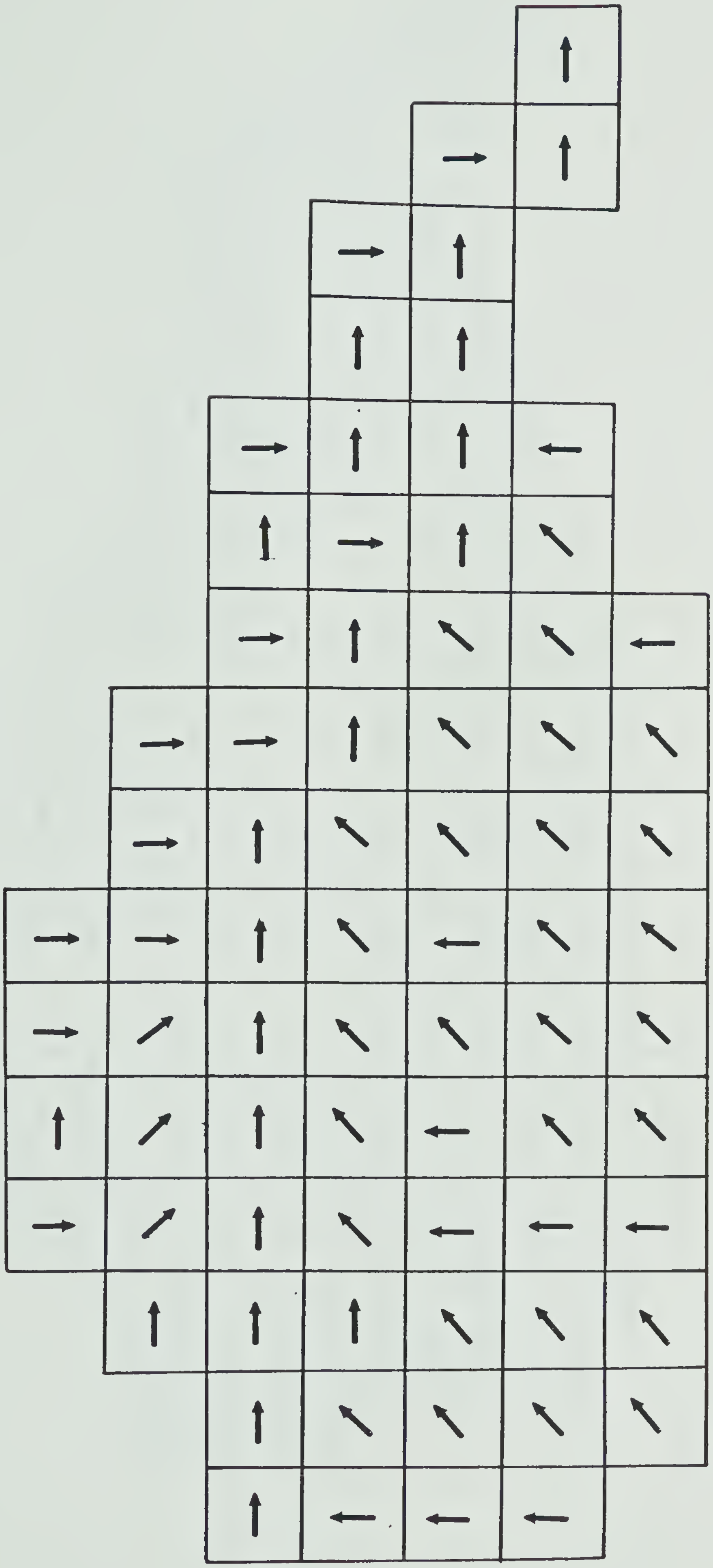


Figure 15a 200 m x 200 m grid system

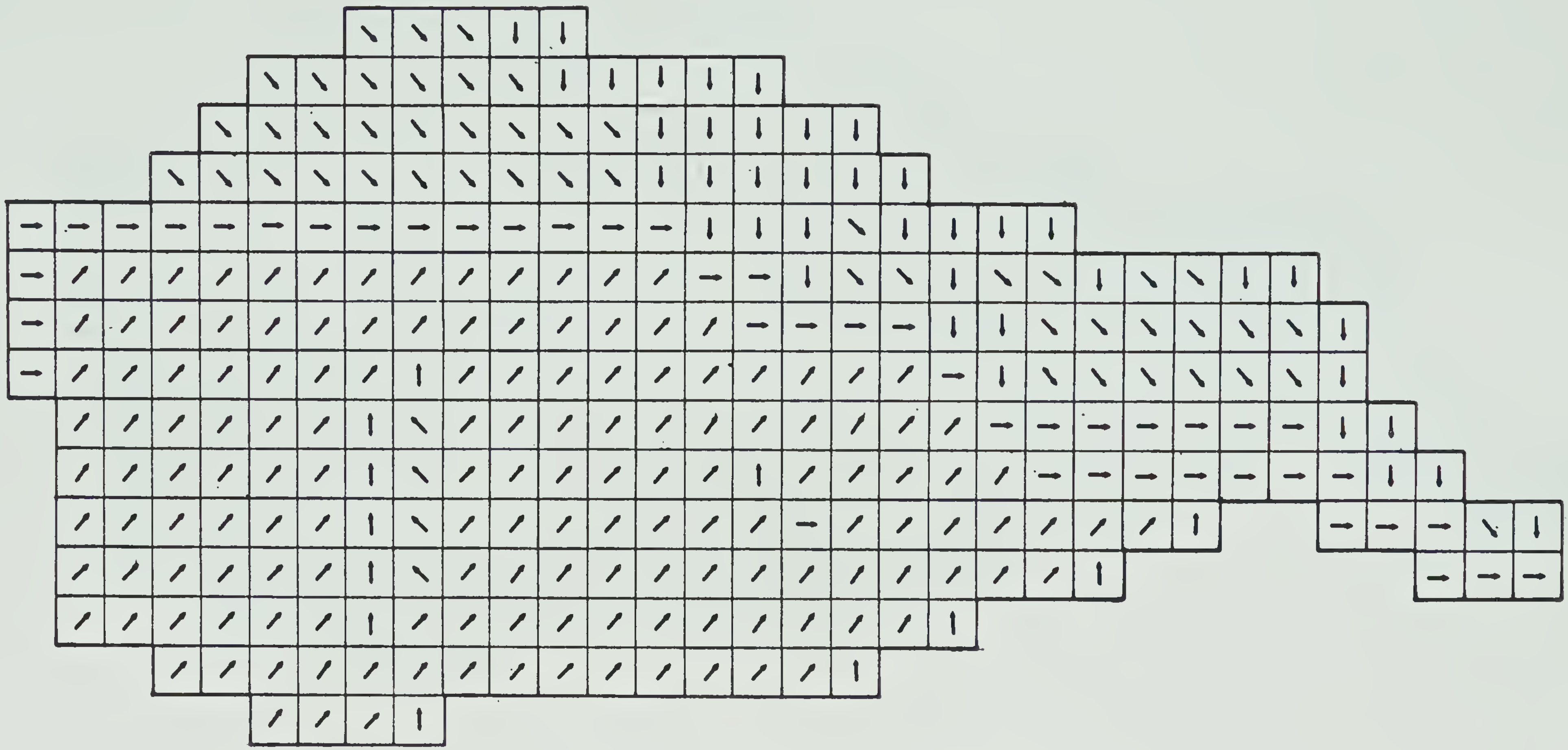


Figure 15b 100 m x 100 m grid system

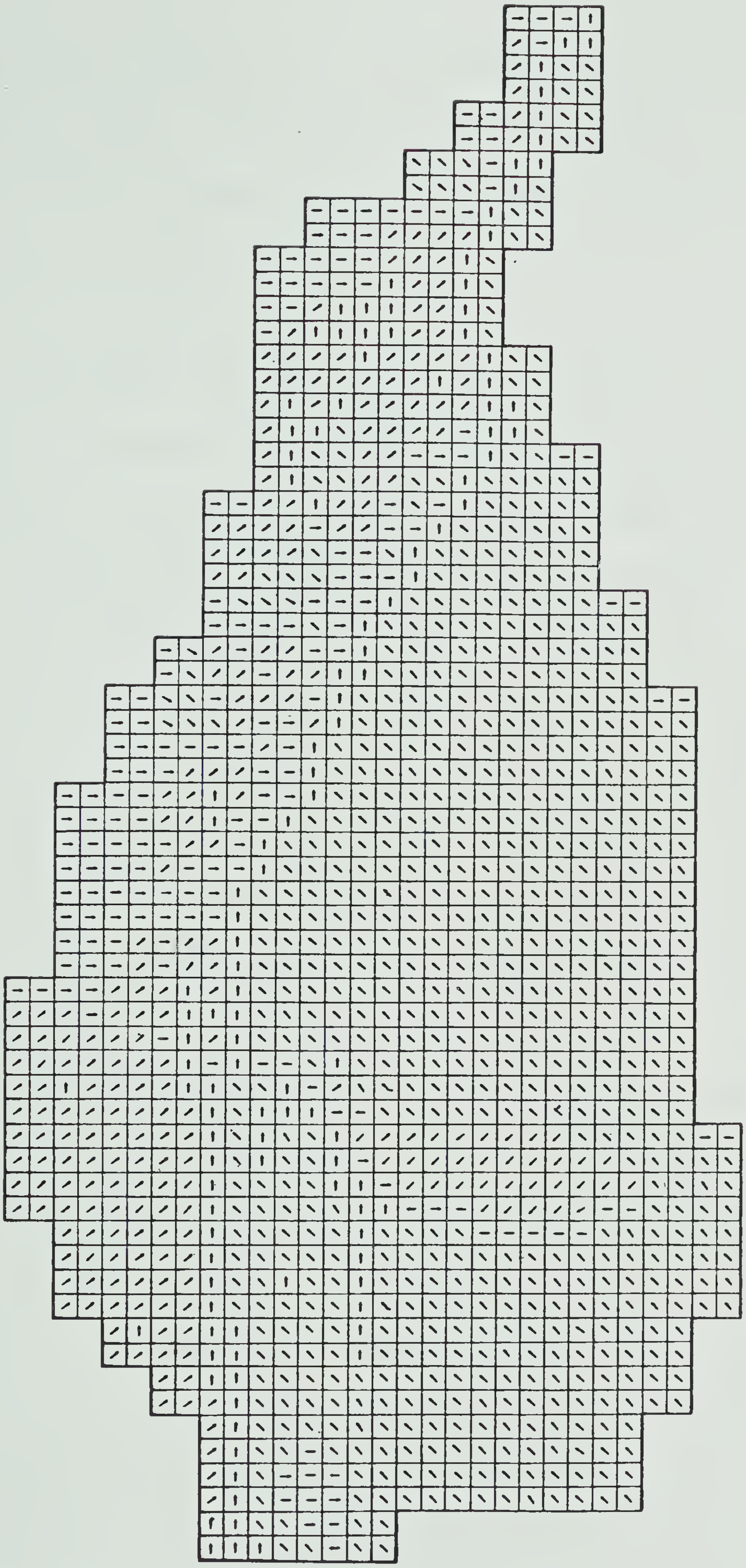


Figure 15c 50 m x 50 m grid system

systems from least to most intense detail with the flow arrows shown on each. Increasingly more topographic detail can be seen to be derived through the use of smaller elements.

8.2.2.2 Data Base

Streamflow is measured by a 120° V notch weir at the mouth of Jamieson Creek and storm rainfall was averaged from 6 recording precipitation gauges located within the watershed.

Cheng (1975) analyzed discharge records for 41 storms that occurred between 1970 and 1974 on the Jamieson Creek watershed. During this period the rainfall amount per storm event varied from 5 to 330 mm, with the majority of the storm durations ranging from 20 to 60 hours. The proportion of storm rainfall that appeared as streamflow varied from 2.5 to 81%, averaging 44%. Instantaneous peak flows ranged from 100 to approximately 15,000 m^3/h.

Of the 41 storms, seven were chosen for simulation. Table 2 presents information about the seven storms that were to be simulated. These particular storms were chosen to give a range of storm magnitudes for simulation. To gain an indication of the antecedent precipitation conditions for each of the storms, daily precipitation data for the Seymour Falls recording station, 12 kilometers to the south of Jamieson Creek were examined and are presented in Table 3. Information on the timing of the individual storms that were simulated and their antecedent conditions can be obtained using Tables 2 and 3. The data reveal that the antecedent conditions for storms 35, 36 and 37 were the driest

Table 2* Simulation Storm Data

Storm	Date	Antecedent Baseflow	Peak Flow	Precipitation	Discharge
		m^3/h	m^3/h	mm	mm
2	November 10, 1970	435	2831	51	12
4	September 4, 1971	330	3217	38	14
8	October 3, 1971	512	8983	79	49
11	October 24, 1971	773	9217	76	42
35	October 20, 1974	56	397	15	1
36	October 27, 1974	59	2552	36	8
37	November 5, 1974	86	8400	80	32

* Adapted from Cheng (1975)

Table 3 Daily Precipitation for Seymour Falls (mm)

Date	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	Total
<u>1970</u>																																
October				10	3		1	9		3	4						16	31	77	13	36	61	55									319
November					33	12	50	30	4	43	17	2			66	19	10	3					62			14	12		17	15		409
<u>1971</u>																																
August	1												8	3					2	20	34									52	37	157
September	21	5	2	32	3		1	13		32													6	1	2	68	32					218
October		33	78	4						2	1	61		1			5	35	15	16	56	19		49	65				7	1	46	494
<u>1974</u>																																
September					1			93	13																	2		1		8		118
October	34	1							3	9	1								19	1					5	21	38		3	2		137
November				24	34	6	8	30	22	38	58				2	12	53	26	47	62	15	28	121	27				4	9			626

Average Monthly Precipitation (mm):

August	88
September	185
October	430
November	488

⊢ ⊣ Storm # (see Table 2)

while those for storms 8 and 11 were the wettest.

The six precipitation measuring gauges, as shown in Figure 16, are fairly well distributed throughout the watershed. One gauge is located at an elevation of 427 m, two are at an elevation of 640 m, two are at 853 m, and one is at 1189 m. To check the possible distributed nature of the storms being simulated, plots of cumulative precipitation versus time were made for storms 35, 36 and 37. The cumulative totals differed insignificantly, suggesting that an average value of precipitation could be used over the entire watershed with confidence. A plot of cumulative precipitation versus gauge elevation also revealed insignificant differences.

8.2.2.3 Parameter Optimization

Storm 11, which occurred on October 24 - 25, 1971, was chosen as the parameter optimization storm because antecedent conditions for this storm suggested that moisture conditions on the watershed were most likely very near field capacity at the start of the storm; thereby reducing the uncertainty of initial moisture conditions. The parameter optimizing simulations were conducted on an hourly basis using the 75 element grid system. The five most sensitive model parameters (saturation, field capacity, conductivity coefficient, profile depth, and drainage coefficient) were varied over their expected range and the effects on the outflow hydrograph were noted. Figures 17a to 17e inclusive present the results of the optimization trials for the various parameters and show the relative effects on the

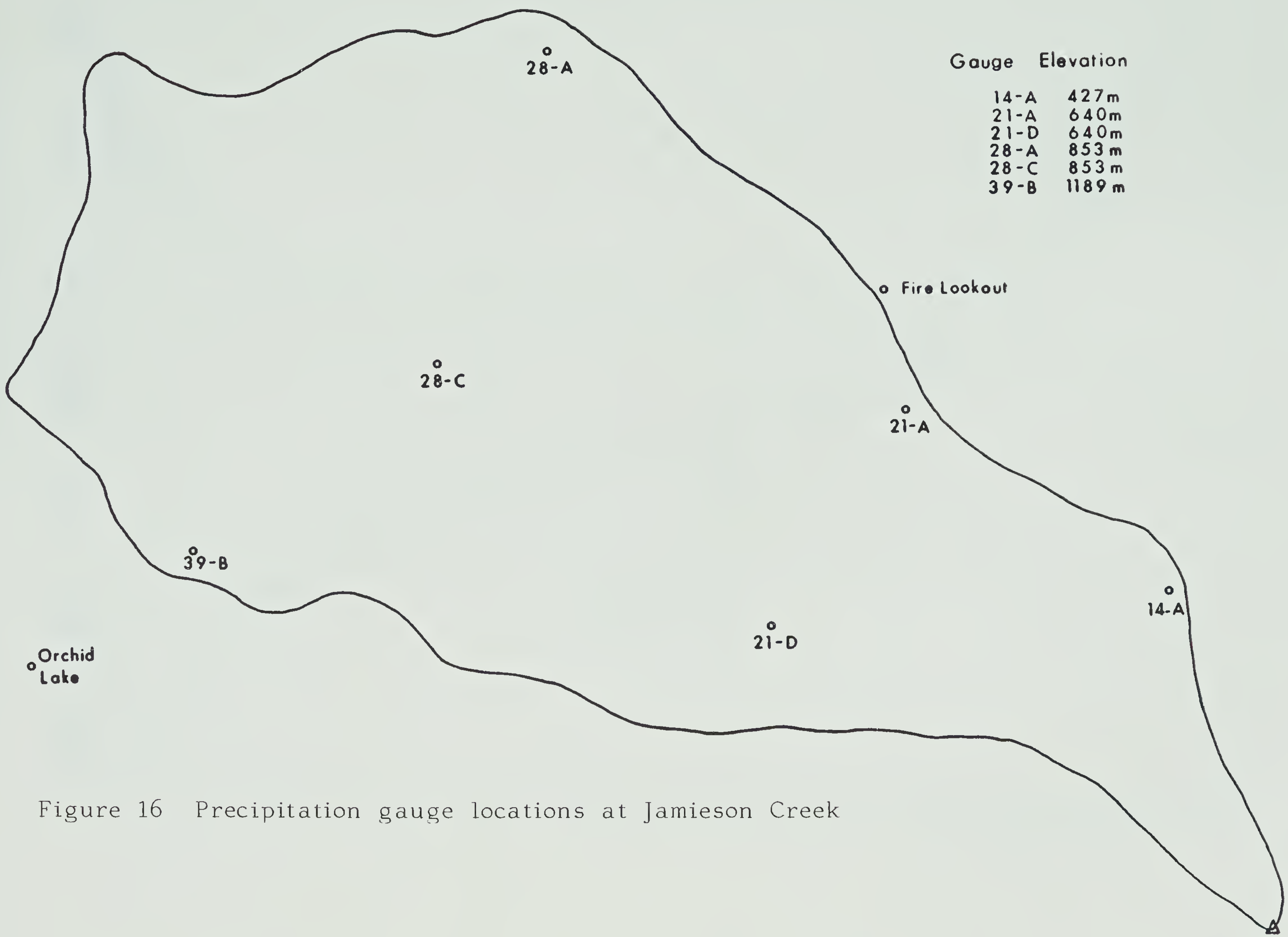


Figure 16 Precipitation gauge locations at Jamieson Creek

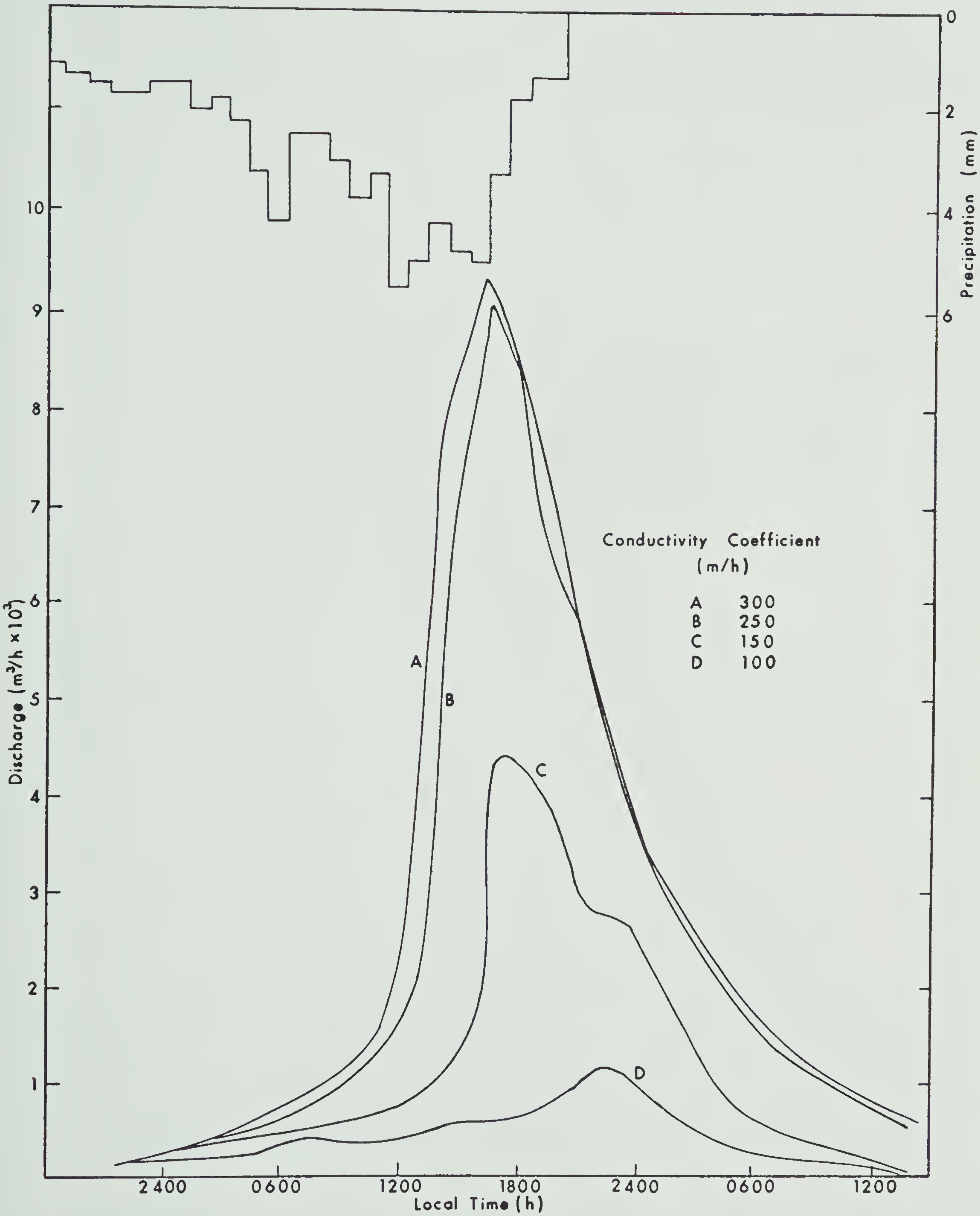


Figure 17a Parameter sensitivity tests
Variable conductivity coefficient

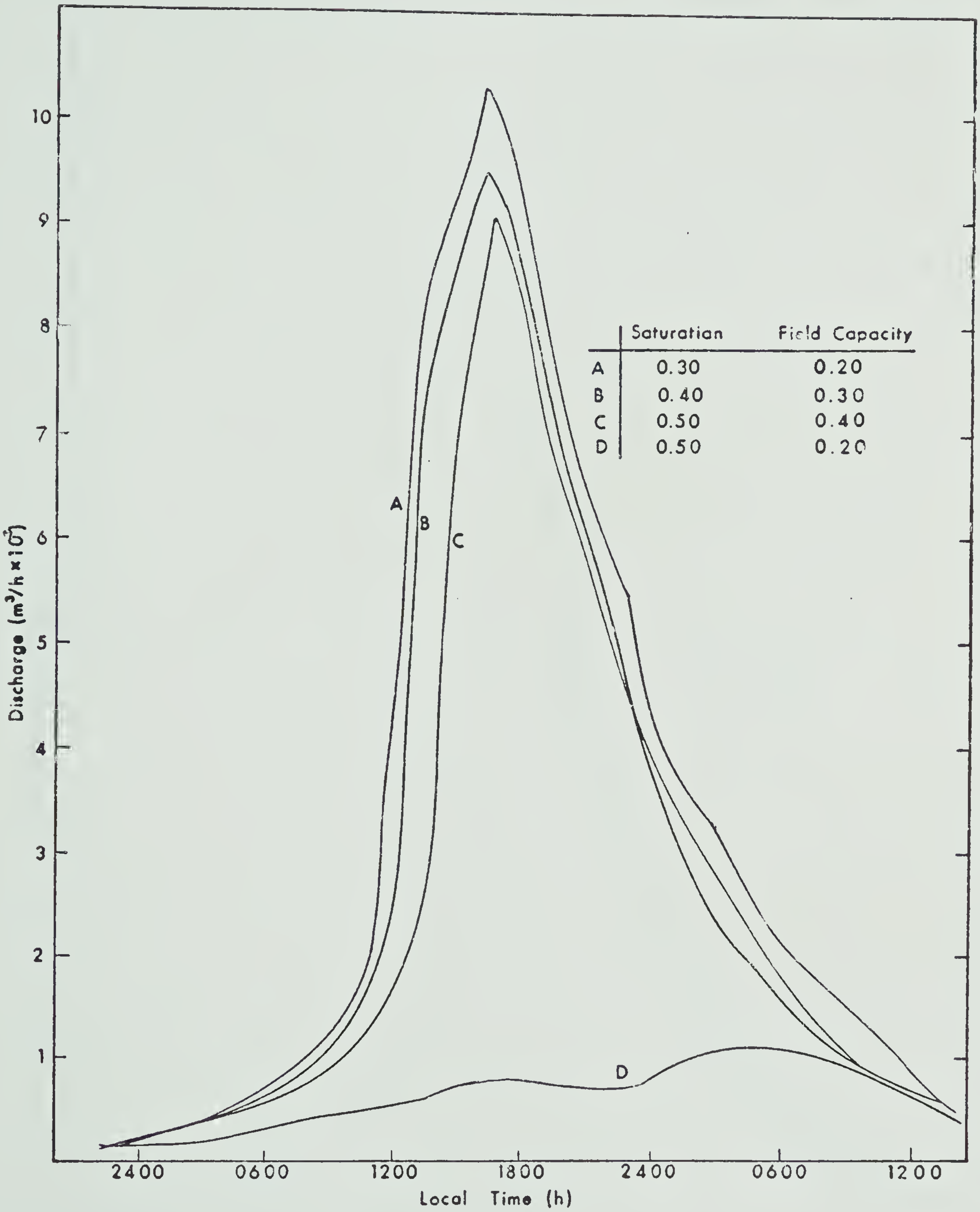


Figure 17b Parameter sensitivity tests
Variable soil moisture storage coefficients

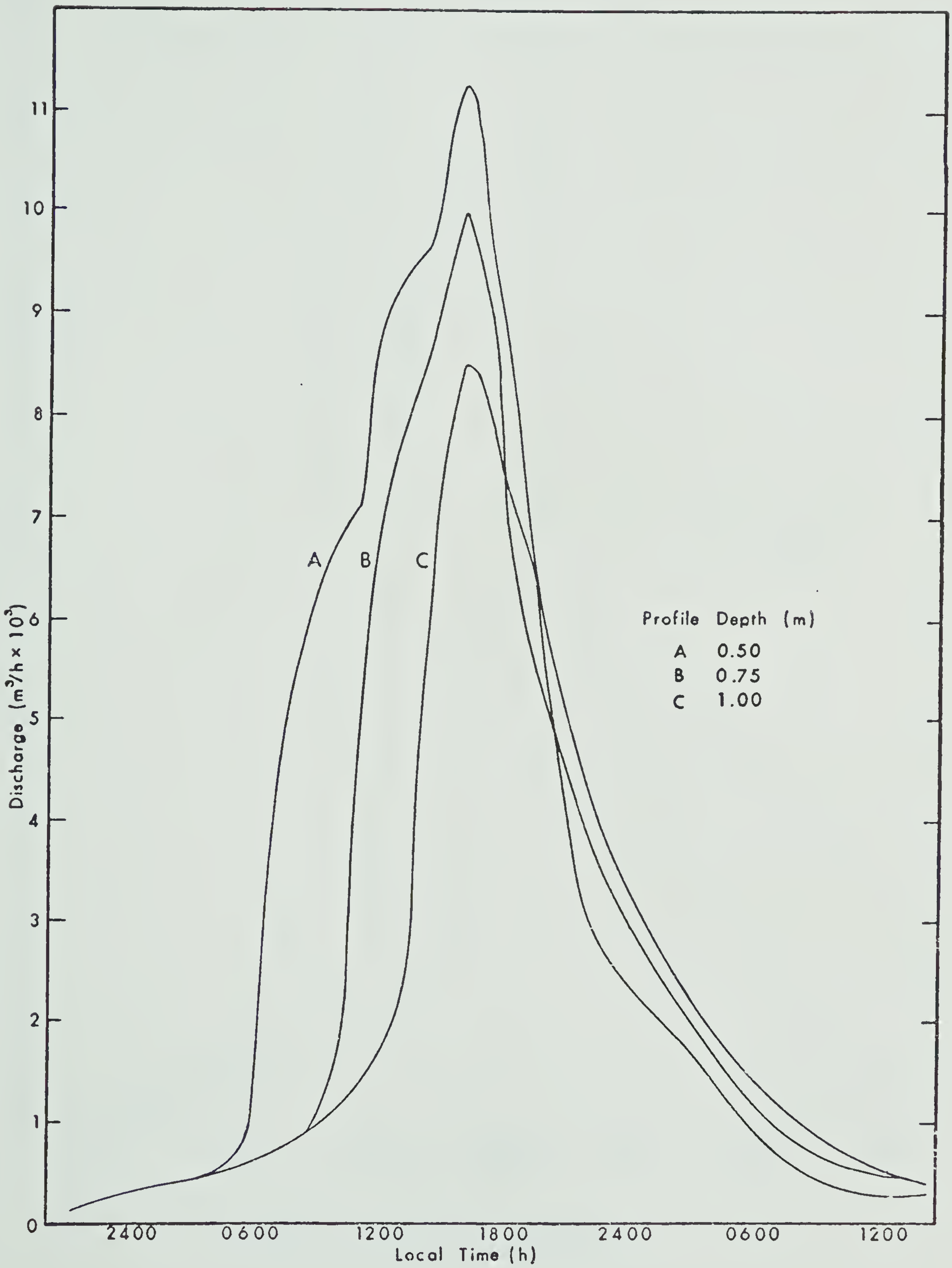


Figure 17c Parameter sensitivity tests
Variable soil profile depth

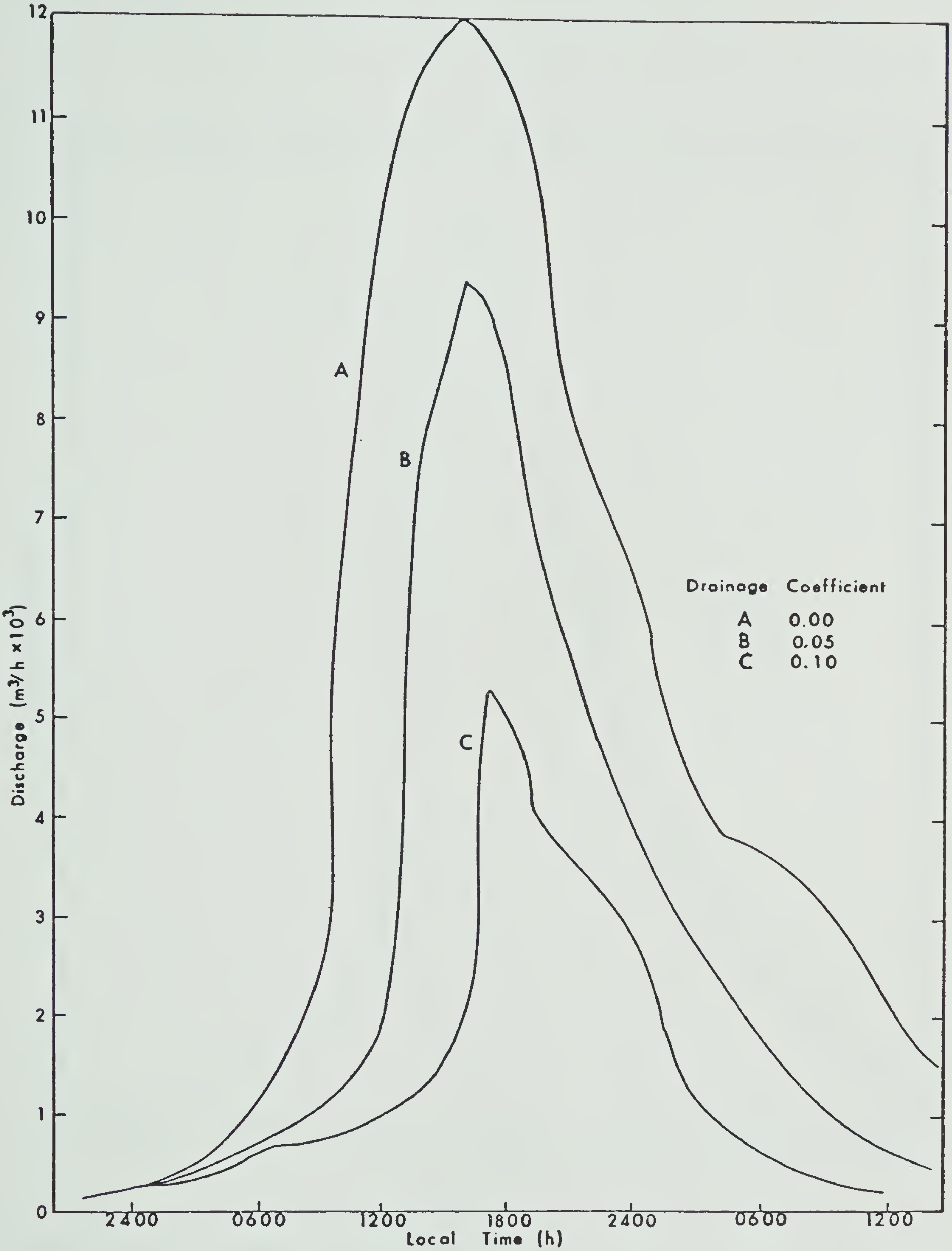


Figure 17d Parameter sensitivity tests
Variable drainage coefficient

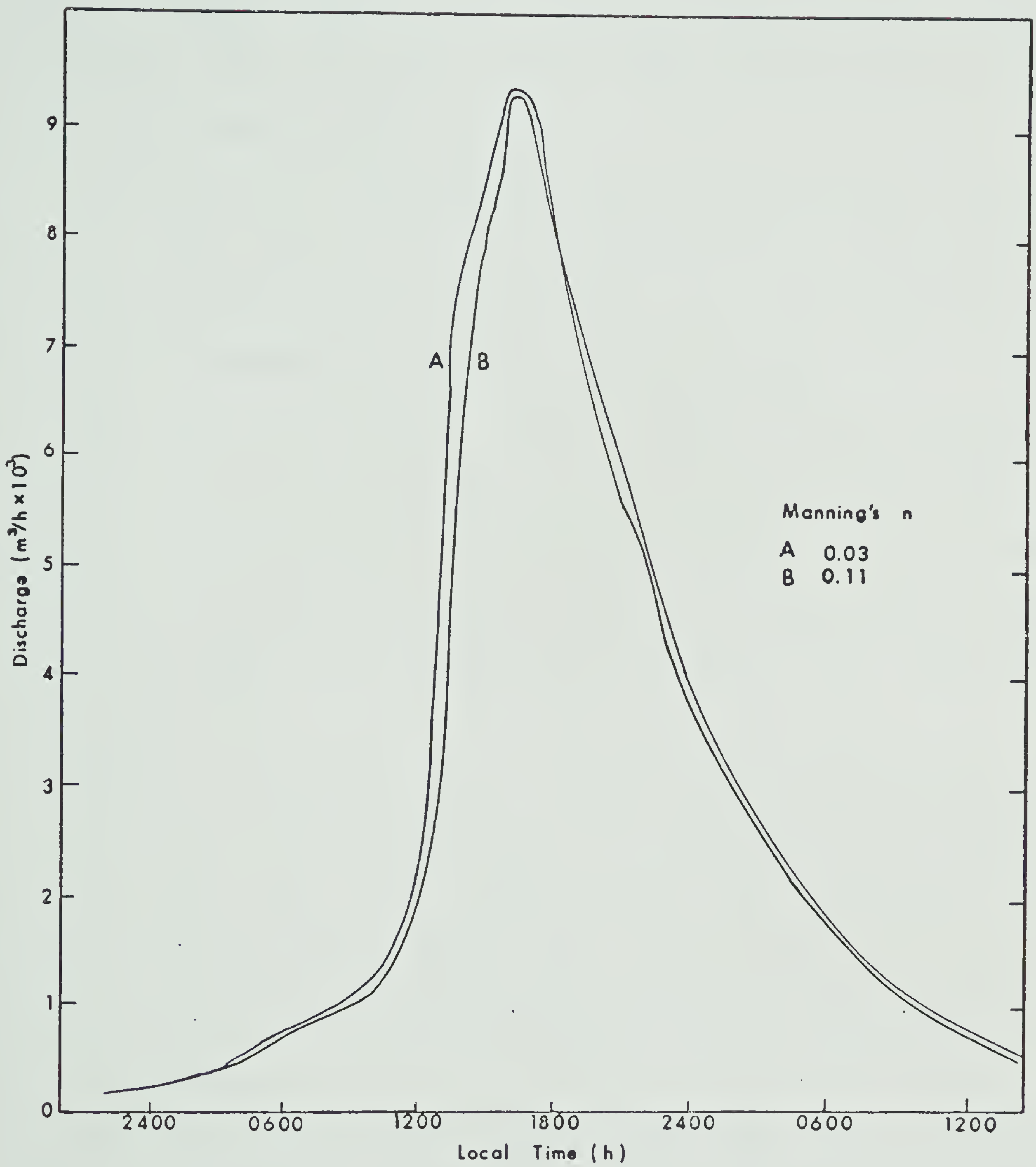


Figure 17e Parameter sensitivity tests
Variable Manning's n

discharge hydrograph due to changes in them. Note that saturation and field capacity values are presented in combination and that Manning's n is included in the optimization. These figures clearly demonstrate the sensitivity of the model parameters. The conclusions drawn after numerous optimization trials for a peak of 9200 m^3/h were:

(1) Use of a saturated coefficient of 0.60 caused a peak that was too low in magnitude and too late in time. Reduction of the coefficient to 0.50 improved the agreement for both magnitude and timing.

(2) Use of a field capacity value of 0.40 provided reasonable agreement between simulated and actual peak discharge. Increasing the field capacity from 0.40 to 0.45 raised the peak to an unacceptably high value and decreased the recession value after 35 hours.

(3) A drainage coefficient of zero caused a peak discharge that was too high. Use of a coefficient of 0.10 lowered the peak to a value below that measured and significantly reduced the value after 35 hours. A value of 0.05 was subsequently found to give the best results.

(4) Reduction of the profile depth from 1.0 to 0.50 m increased the magnitude of the flows on the rising limb, increased the magnitude of the peak, and lowered the values on the recession limb.

(5) A value for the conductivity coefficient of 250 m/h

was required to provide a reasonable peak value as well as a recession value after 35 hours.

Conclusion: The optimum set of parameter values for storm 11 are saturation = 0.50, field capacity = 0.40, drainage coefficient = 0.05, depth of profile = 1.0 m, and conductivity coefficient = 250 m/h.

8.2.2.3.1 Correlation of Optimized and Measured Parameters

As a check to see if the optimized parameter values correspond to literature reported values, estimates of the parameter values for the soil moisture storages coefficients were obtained from Cheng (1975). Figure 18 is a laboratory-derived water characteristic curve for three soil layers of a Strachan gravelly sandy loam from Jamieson Creek. The volumetric moisture content at zero tension corresponds to the saturation moisture content while the volumetic moisture content at 100 cm tension corresponds to field capacity. Figure 18 reveals that the saturation occurs at a moisture content of 0.50 while field capacity occurs at 0.20. The optimized saturation value is identical to the measured one but the optimized field capacity is considerably higher than the 'measured' value. The agreement for saturation is extremely encouraging, because the coefficient has definite, measurable physical significance. The lack of agreement for field capacity can possibly be explained as a matter of interpretation. Field capacity is supposedly the moisture content that results when the free drainage of water ceases. However, a sharp break in the moisture content versus

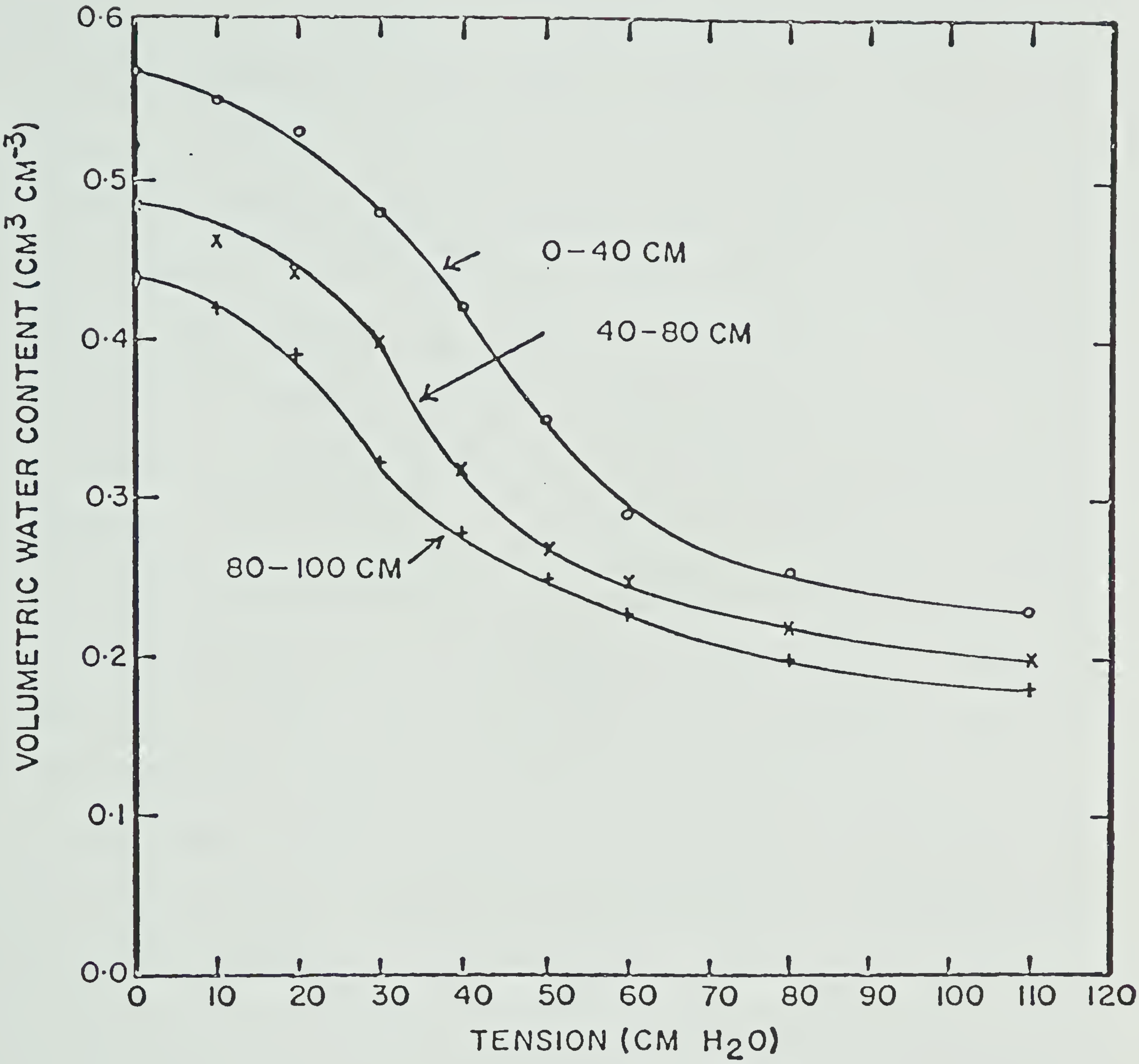


Figure 18 Water retention characteristics of Strachan gravelly sandy loam (Cheng, 1975)

time curve for any soil as suggested by this concept, does not exist. Therefore, field capacity, since it can be a useful concept, has to be defined otherwise to facilitate laboratory measurement. In the case of the sandy loams of Jamieson Creek, it would be defined as the moisture content that would exist at a tension of 100 cm. Since this parameter is arbitrarily set, the disagreement between model optimized and 'measured' values of field capacity is not surprising. The higher optimized value suggests that actually less water is very active in subsurface flow than the concept of field capacity would indicate. Figure 17b displays an extremely low peak and very flat hydrograph if saturation is set at 0.50 and field capacity at 0.20, as measured by the laboratory techniques. This suggests that optimized values for field capacity may not correspond to the laboratory measured values for it. The same conclusion would probably apply to wilting point since similar problems of interpretation may exist.

In view of the discussion concerning the reliability of laboratory measured hydraulic conductivities, a comparison of laboratory measured and optimized values for hydraulic conductivity is meaningless. A soil profile depth of 1.0 m seems quite reasonable based on information from Cheng (1975) and Willington (1971). The drainage coefficient as defined here is extremely difficult to measure in the field and thus the value of 0.05 will be accepted.

8.2.2.3.2 Effect of Storm Selection on Parameter Optimization

To ensure that the parameter optimization just described was not storm selection dependent, the entire optimization

procedure was repeated for storm 2. The optimized values for saturation, field capacity, profile depth and drainage coefficient based on this storm were identical to those optimized for storm 11. The conductivity coefficient showed a minor change in optimized value with 200 m/h giving the best results. This difference in optimized values is not considered to be significant. The fact that the conductivity coefficient should be changed slightly should come as no surprise since this parameter is the most sensitive. The optimization results reveal that the optimization procedure was not storm selection dependent.

8.2.2.3.3 Effect of Grid Size on Parameter Optimization

To evaluate the effect that grid size might have upon the optimized parameters, the entire optimization procedure was now repeated for the watershed using the 300 element (100 m X 100 m) grid. Conclusions about the effects on optimized parameters were found to be identical for those discussed above for the 75 element grid, for storms 2 and 11, with one notable exception. A conductivity coefficient of 250 m/h produced a peak discharge higher than the recorded one. Reduction of the coefficient to 150 m/h produced closer agreement between the simulated and the actual peaks. This reduction may have occurred as a result of the increase in topographic detail obtained through the use of more grid elements, allowing for the formation of more saturated elements. This would cause a larger portion of the outflow to occur as overland flow.

Further optimization trials were conducted for the 1200

element (50 m X 50 m) grid. All parameters were optimized at values identical to those for the two coarser grids, with the exception of the conductivity coefficient, which required a decrease to 100 m/h, from 150 m/h. One possible explanation for this decrease in conductivity coefficient is that the increased detail in topographic features has affected the flow paths and thus the conductivity coefficient, in a fashion similar to the one discussed previously for the 300 element grid.

8.2.2.4 Distributed Soil Properties

The assumption of uniform soil properties throughout the watershed is an unrealistic one. In nature, homogeneity is the exception rather than the rule. Soil profile depth could be expected to vary along any given slope, with the deepest profile at the toe of the slope and the shallowest at the ridge. As well, drainage coefficients might also be expected to be dependent on slope position.

In order to implement the more natural variation in these soil properties, slope cross-sections were drawn along grid columns of the watershed. Low-lying elements were assessed soil profile depths of 1.0 m, midslope elements 0.75 m and near ridge elements 0.50 m. Low-lying elements were assessed drainage coefficients of 0.00 while and all other elements 0.05. Simulation for storm 11 was repeated using these distributed soil parameters for the 300 element grid with all other parameters held at their optimized values. Results showed that all simulated discharges were now higher, with the peak approximately 25% higher.

Further simulations revealed that the conductivity coefficient could be reduced to 50 m/h to match the measured discharge peak and simulated one. Again the decrease in the optimized conductivity coefficient should come as no surprise since the decrease in profile depths at higher elevations means reduced water holding capacities in these areas. The result is higher interflow volumes to the low lying areas which become saturated more quickly resulting in transmission of flow by overland routes rather than by subsurface routes. The actual shape of the hydrograph remained very similar to the one for the 75 element grid.

8.2.2.5 Simulation Results

Simulation results for all seven storms for all three grid systems are presented in Figures 19a to 19g inclusive. Note that only the 300 element grid has distributed properties of soil profile and drainage coefficient. Recall that the 75 element simulations required a conductivity coefficient of 250 m/h, the 300 element ones 50 m/h and the 1200 elements ones 100 m/h. Peaks for the 300 element grid are generally slightly higher than those for either the 75 or the 1200 element grid. The 75 element simulations show rather subdued peaks for storms 4 and 36. The 1200 element simulation peaks appear to be delayed a few hours in several instances.

Examination of the simulation results for storms 8 and 11 suggest that the 300 element grid provides the best results of the three options, with the excellent agreement on the recession

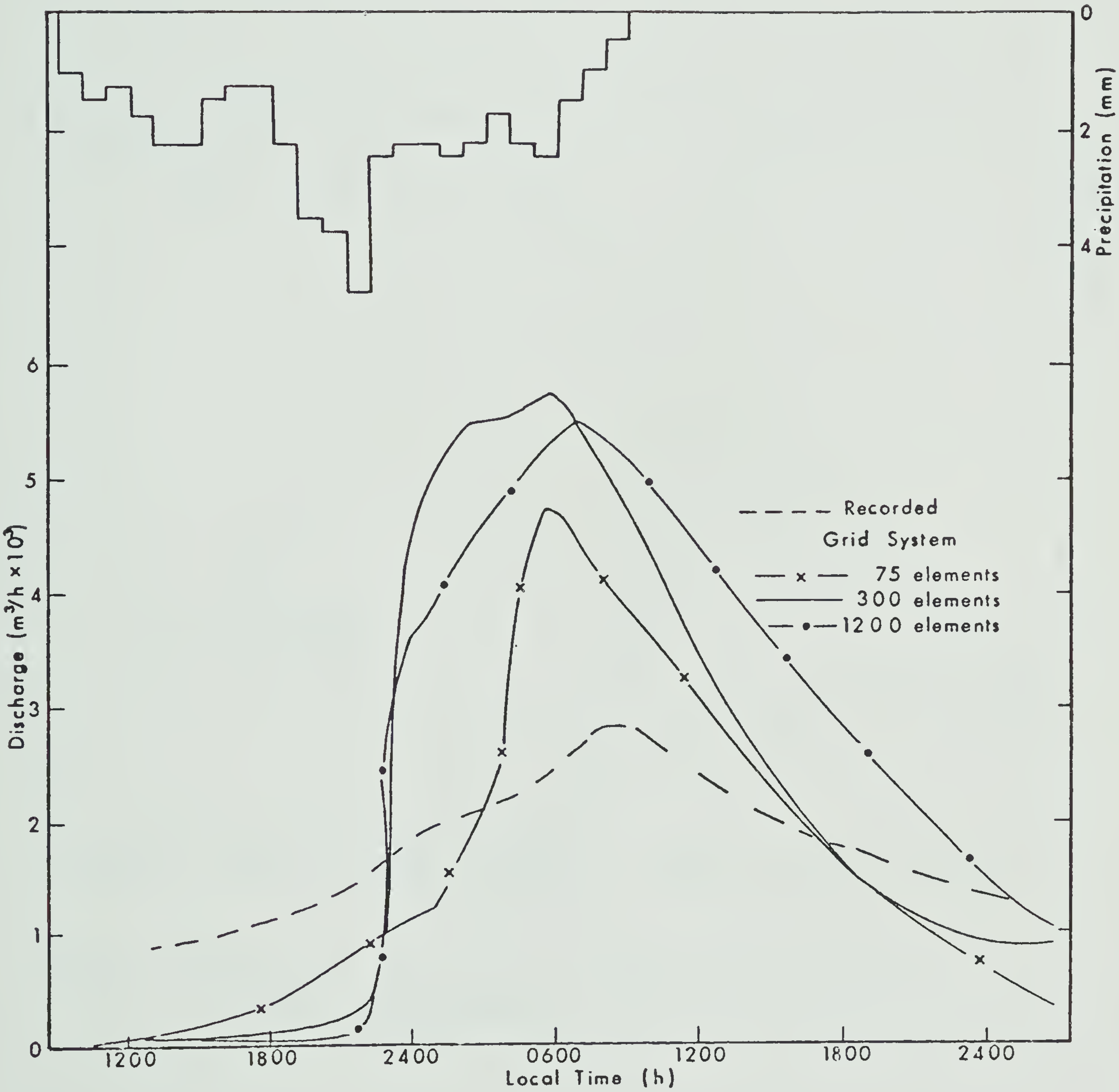


Figure 19a Effect of grid size on storm hydrographs
Storm 2 simulations

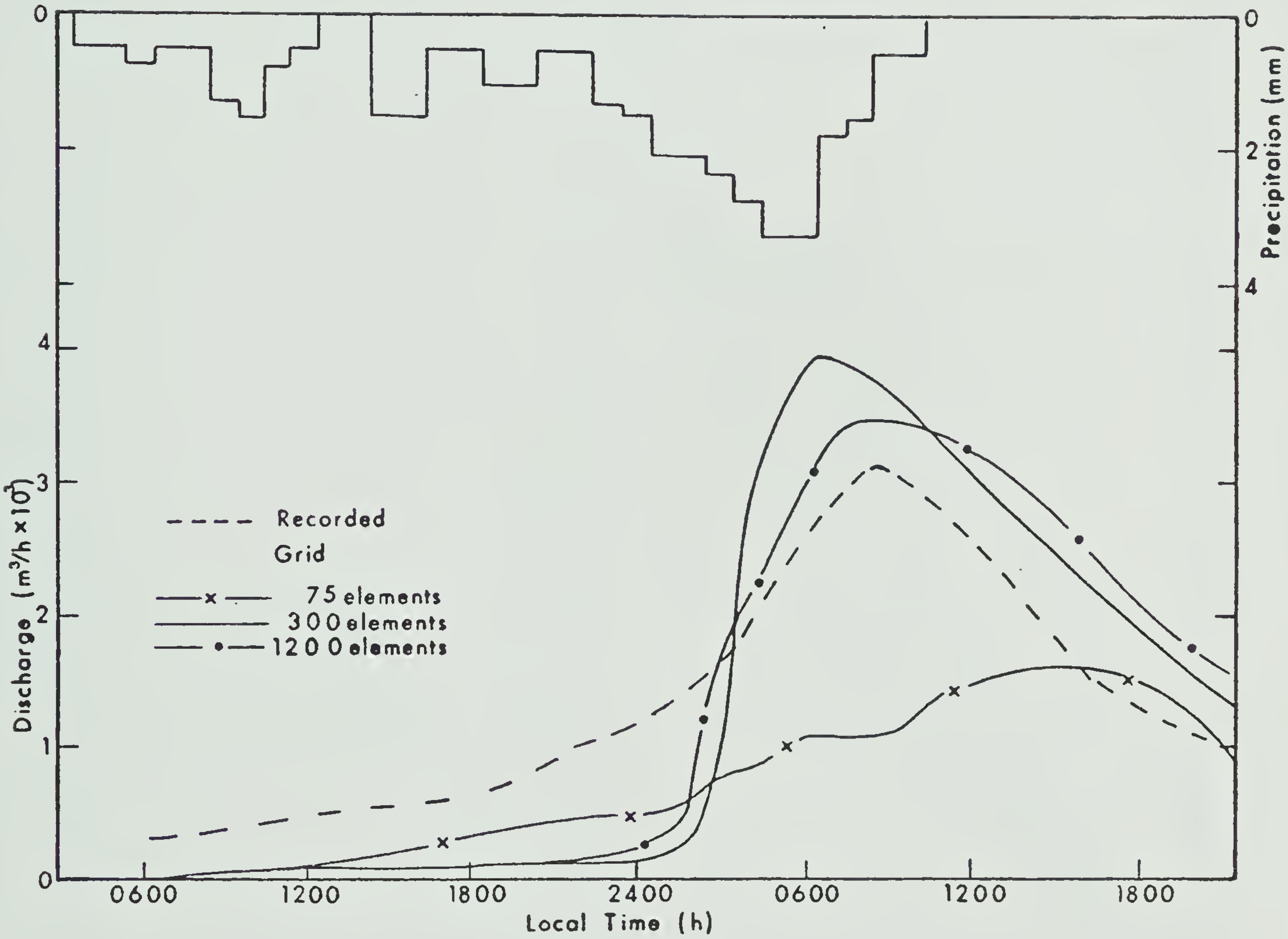


Figure 19b Effect of grid size on storm hydrographs
Storm 4 simulations

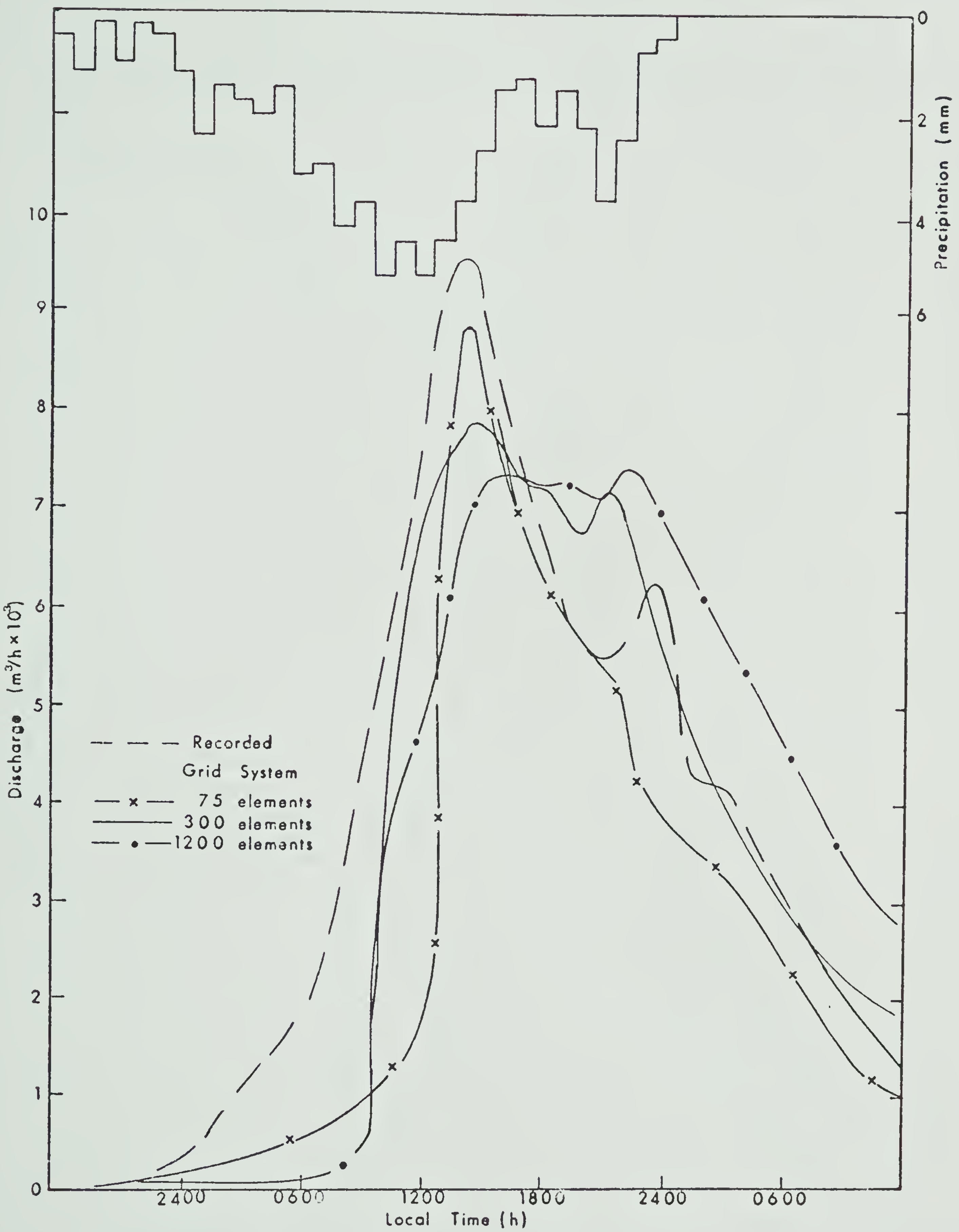


Figure 19c Effect of grid size on storm hydrographs
Storm 8 simulations

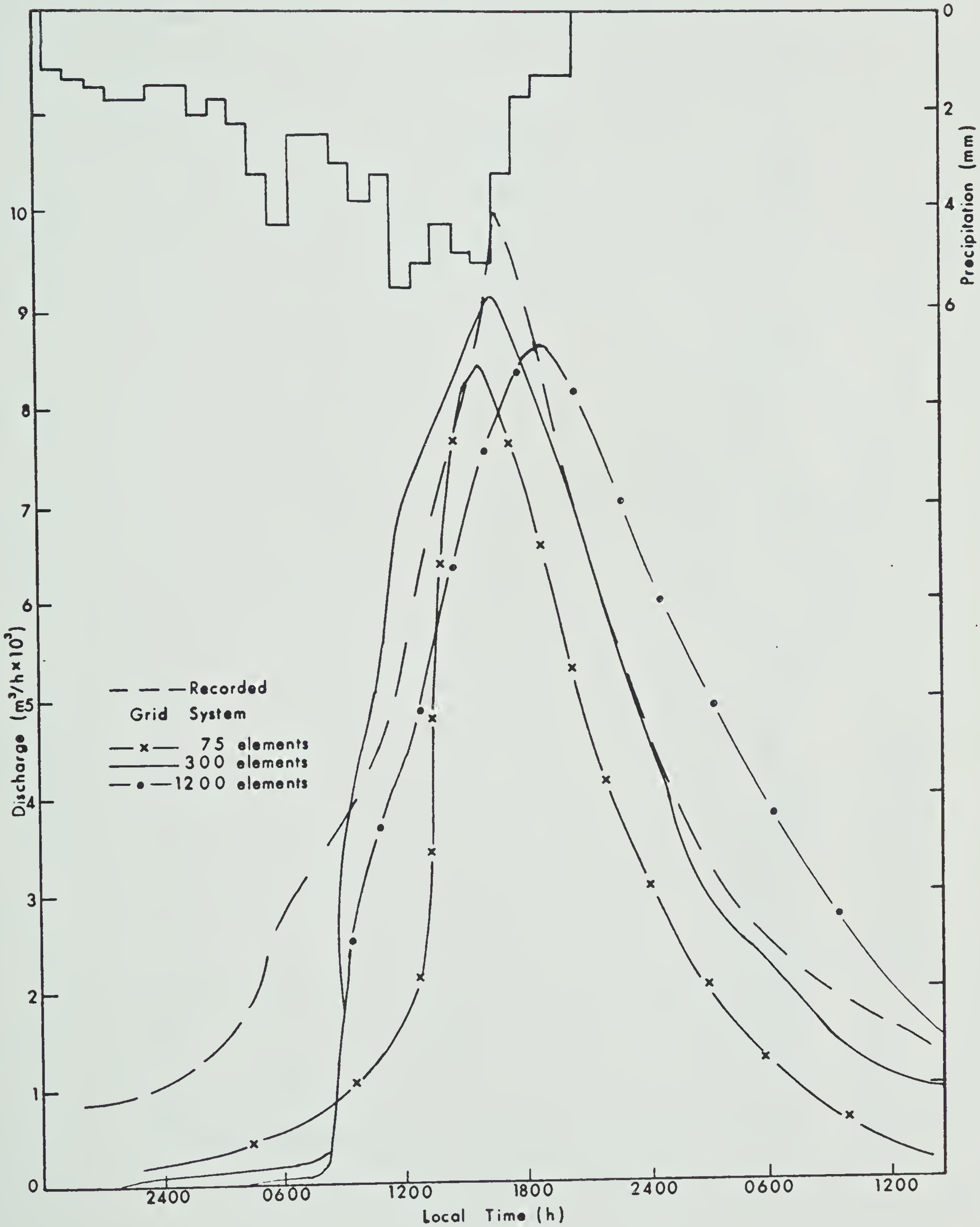


Figure 19d Effect of grid size on storm hydrographs
Storm 11 simulations

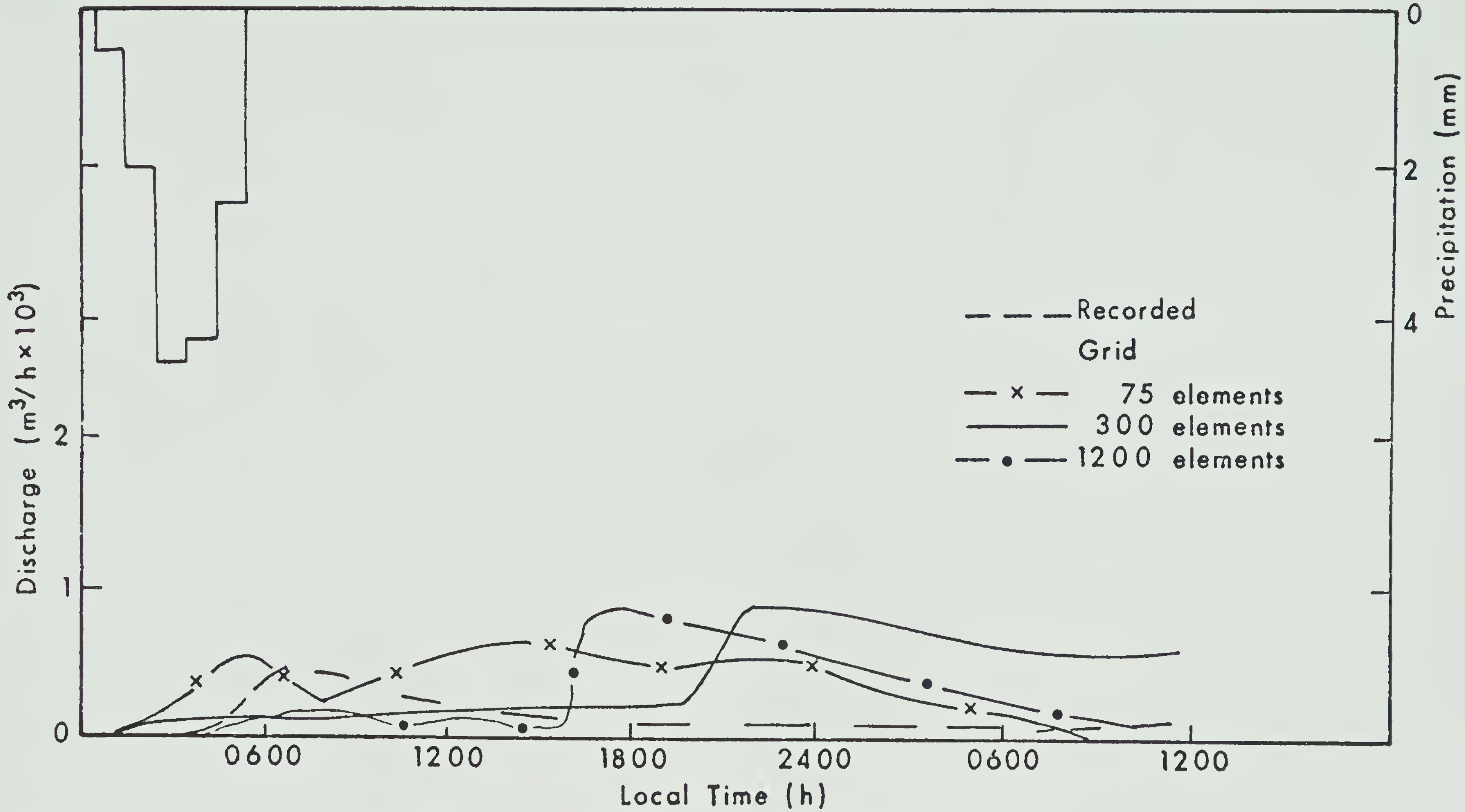


Figure 19e Effect of grid size on storm hydrographs
Storm 35 simulations

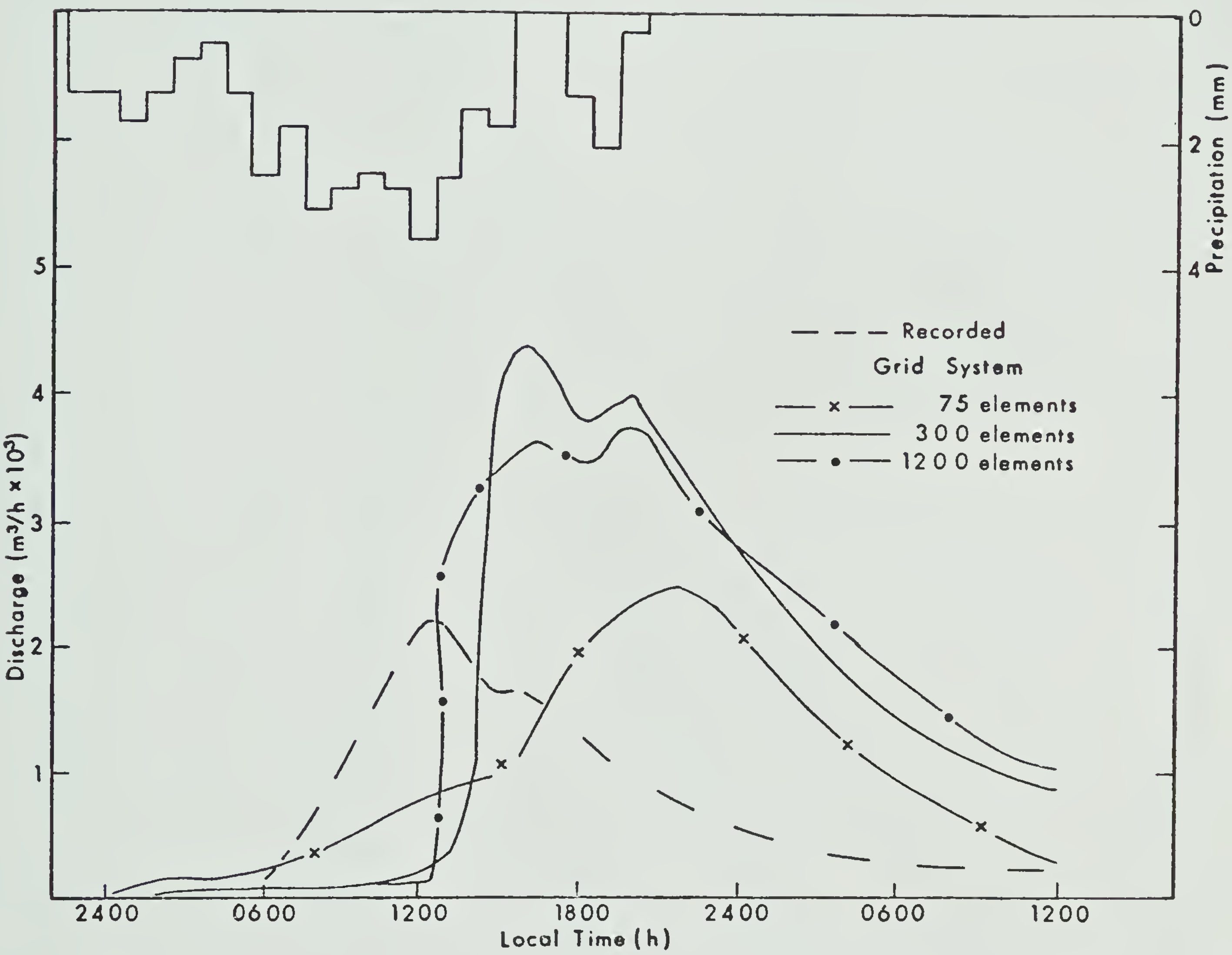


Figure 19f Effect of grid size on storm hydrographs
Storm 36 simulations

Figure 19g Effect of grid size on storm hydrographs
Storm 37 simulations

limb being most noticeable. One must recall that the 1200 element grid requires considerably more topographic data input and storage space than did the 300 element grid; simulation results do not indicate that the extra effort in using more elements is justified. Consequently, the 300 element grid will be used for all further investigations.

8.2.2.6 Initial Conditions

One very noticeable feature about the comparison between the simulated and actual discharge values is the discrepancy during the first half of the rising limb of the hydrograph. The recorded hydrograph shows a gradual increase in discharge while the simulated one remains at a low value and then rises rather abruptly, giving the simulated hydrograph a rather unnatural appearance. This, it was felt, was due to the assumption of uniform moisture throughout the basin; required are distributed moisture contents throughout the watershed.

To examine the effects of initial moisture conditions on the simulated discharge hydrographs, storm 11 was run twice to simulate two storms in succession; the first or 'dummy' storm being used only to initialize moisture conditions for the second storm. The time space between the two storms was varied to examine the effects of variable moisture contents. The second storm was begun when the simulated flow using the 'dummy' storm was at levels of 2241, 794 and 53 m^3/h. These flows occurred 35, 46, and 129 hours after the start of the 'dummy' storm respectively. These different levels of flow would reflect

different moisture conditions in the watershed. Simulation results are shown in Figure 20 and demonstrate the rather dramatic effect of the initial conditions in the watershed upon the discharge hydrograph. However these effects, even for a wide range of initial conditions, persist for only a very short time. In the simulations, by the eighteenth hour of simulation, values of discharge for all three cases are equivalent.

To ensure that the results of the use of a particular 'dummy' storm for initializing purposes were not storm selection dependent, the simulations were repeated now using storm 8 as the 'dummy' storm and storm 11 as the second storm. The resulting discharge hydrograph for the second storm was practically identical to that using storm 11 as the 'dummy' storm. This result suggests that the hydrograph will not depend upon the particular storm used for initializing purposes. This conclusion is very encouraging. Even though the total amounts of precipitation for the two storms 8 and 11 are almost identical, the distribution of precipitation for the two storms is actually quite different (compare Figures 21c and 21d), suggesting the choice of storm for initializing conditions does not limit nor does it significantly affect the simulation of an ensuing storm in an undesirable way. Storms 8 and 11 were chosen initially because the relatively high amount of precipitation in either case should have wetted the watershed throughout and created saturated layers in the low-lying areas.

As a result, the problem of initializing conditions for simulation can be easily solved through the use of a 'dummy'

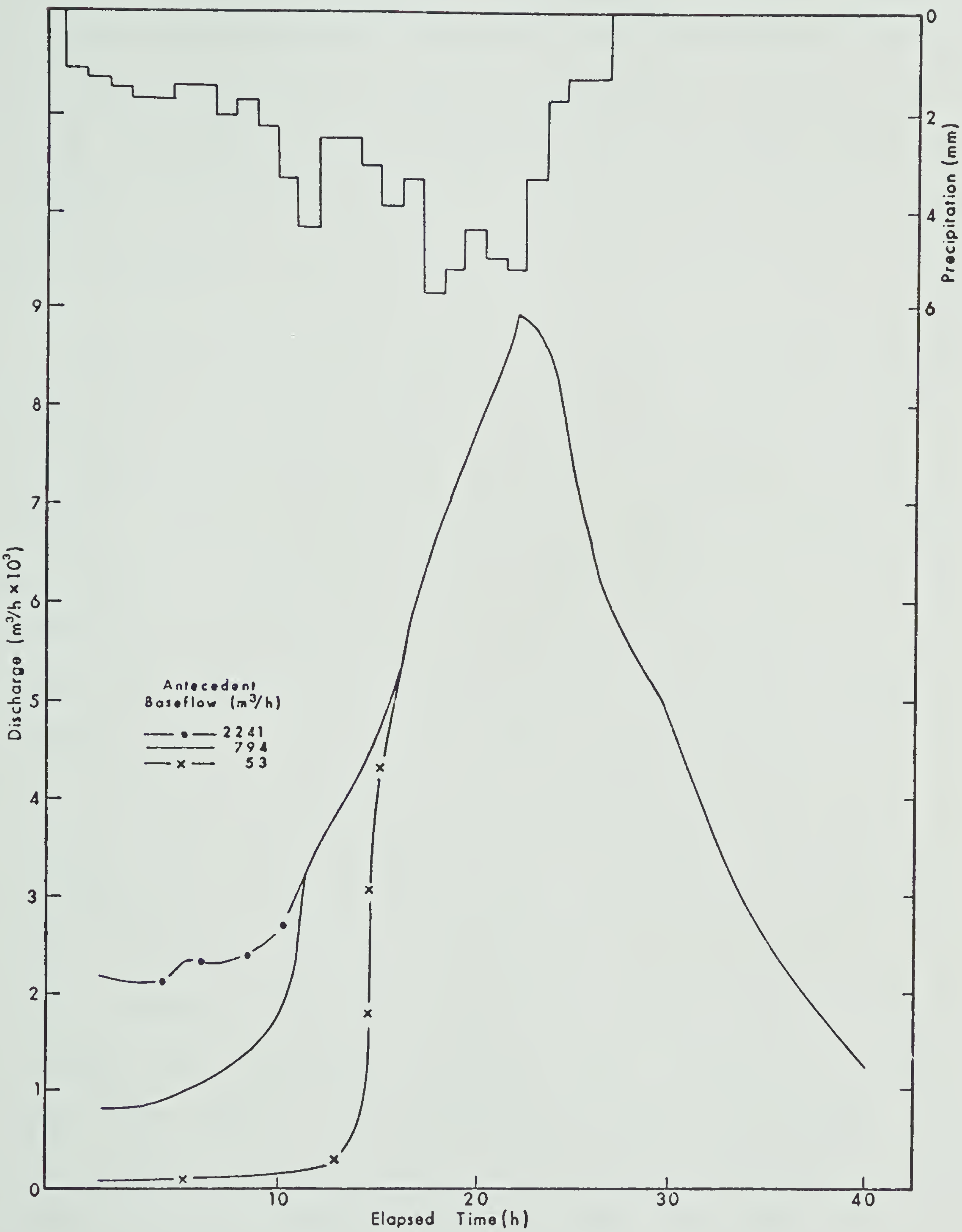


Figure 20 Effect of initial conditions

storm which need only be large enough to have wet the watershed throughout, preferably resulting in a saturated layer in each element sometime during the course of simulation using the 'dummy' storm. This solution is particularly suitable in view of the distributed nature of the model and the large number of elements to be initialized.

8.2.2.7 Goodness of Fit

Simulations were now rerun for all seven simulation storms using storm 11 for the purpose of initializing moisture contents (as opposed to the previous practice of assuming a uniform moisture content of field capacity for all elements) to demonstrate the degree of agreement between the simulated and recorded results for the seven storms. These simulations are shown in Figure 21a through 21g. Storms 2, 4, 8, and 11 all showed a beneficial effect on the rising limb of the hydrograph due to this technique of moisture initialization with storm 11 now showing very good agreement between recorded and simulated discharges on the rising limb. Storms 35, 36 and 37, however, showed negligible changes in hydrograph shape through the use of a 'dummy' storm as opposed to initialization at field capacity. Low antecedent baseflows were recorded for these storms (see Table 2) and apparently the moisture conditions for these storms may have been such that the assumption of a uniform moisture content throughout the watershed was not particularly limiting. Results suggest, however, that agreement between recorded and simulated discharge values on the rising

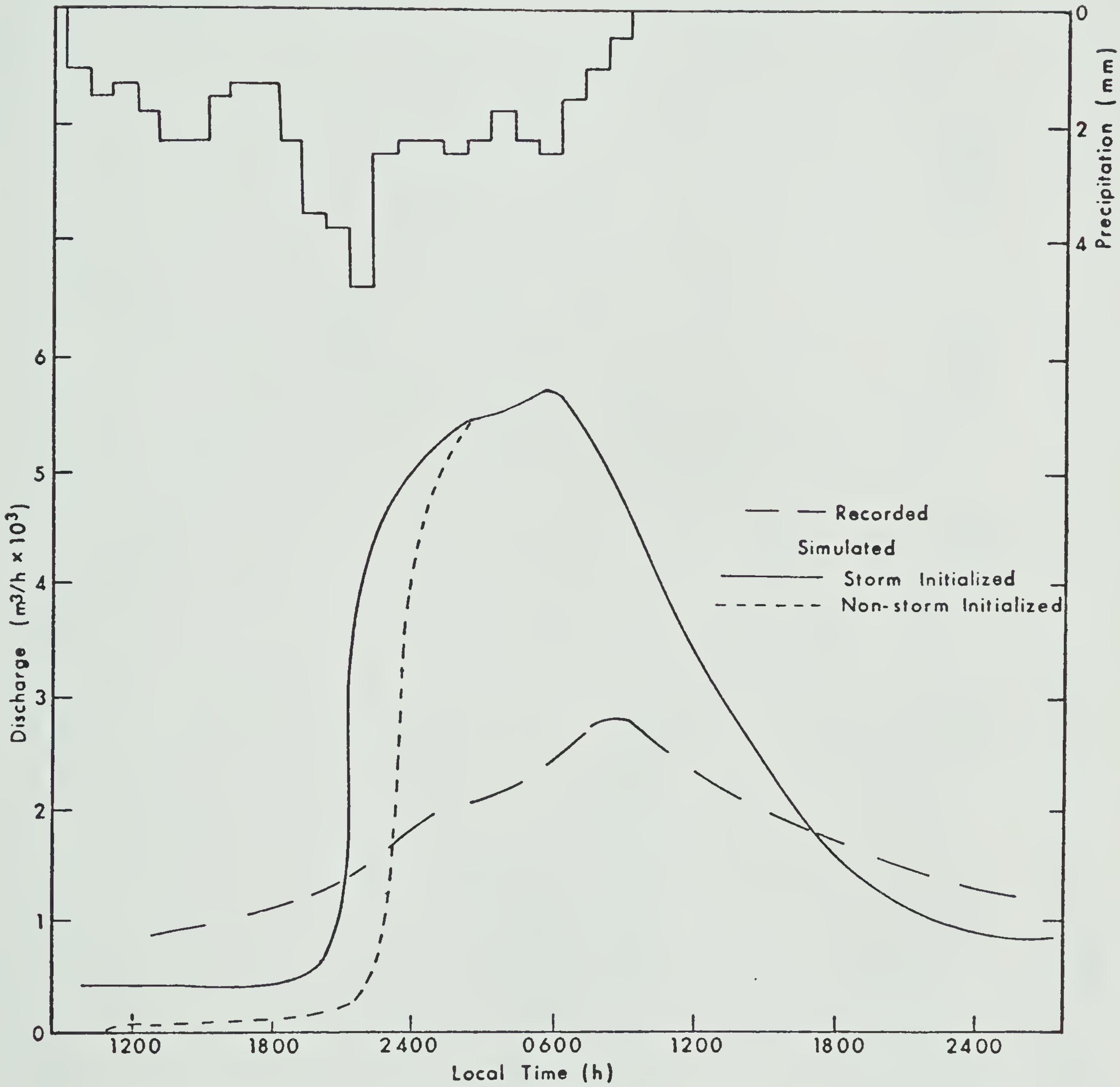


Figure 21a Initializing conditions and goodness of fit
Storm 2

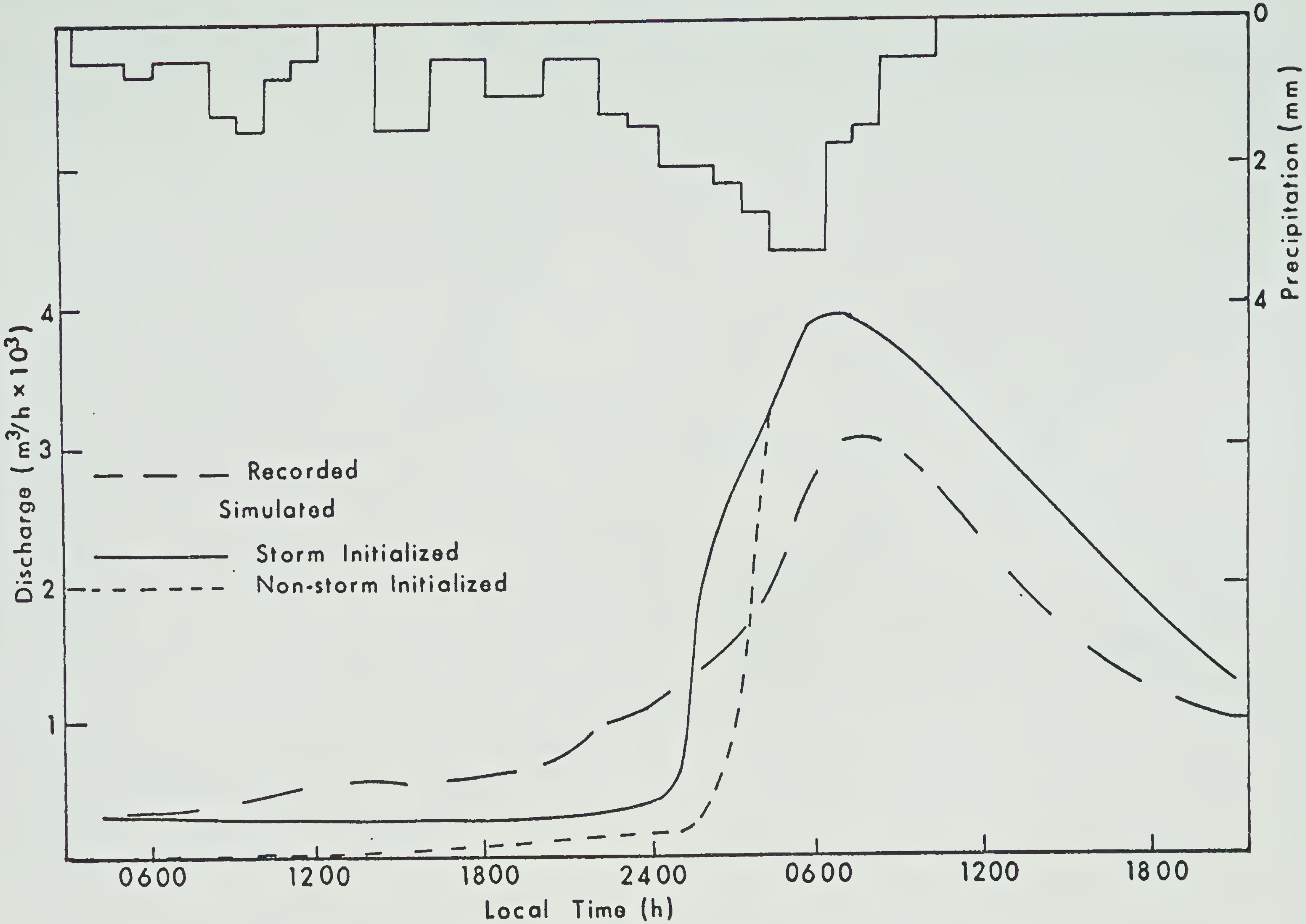


Figure 21b Initializing conditions and goodness of fit
Storm 4

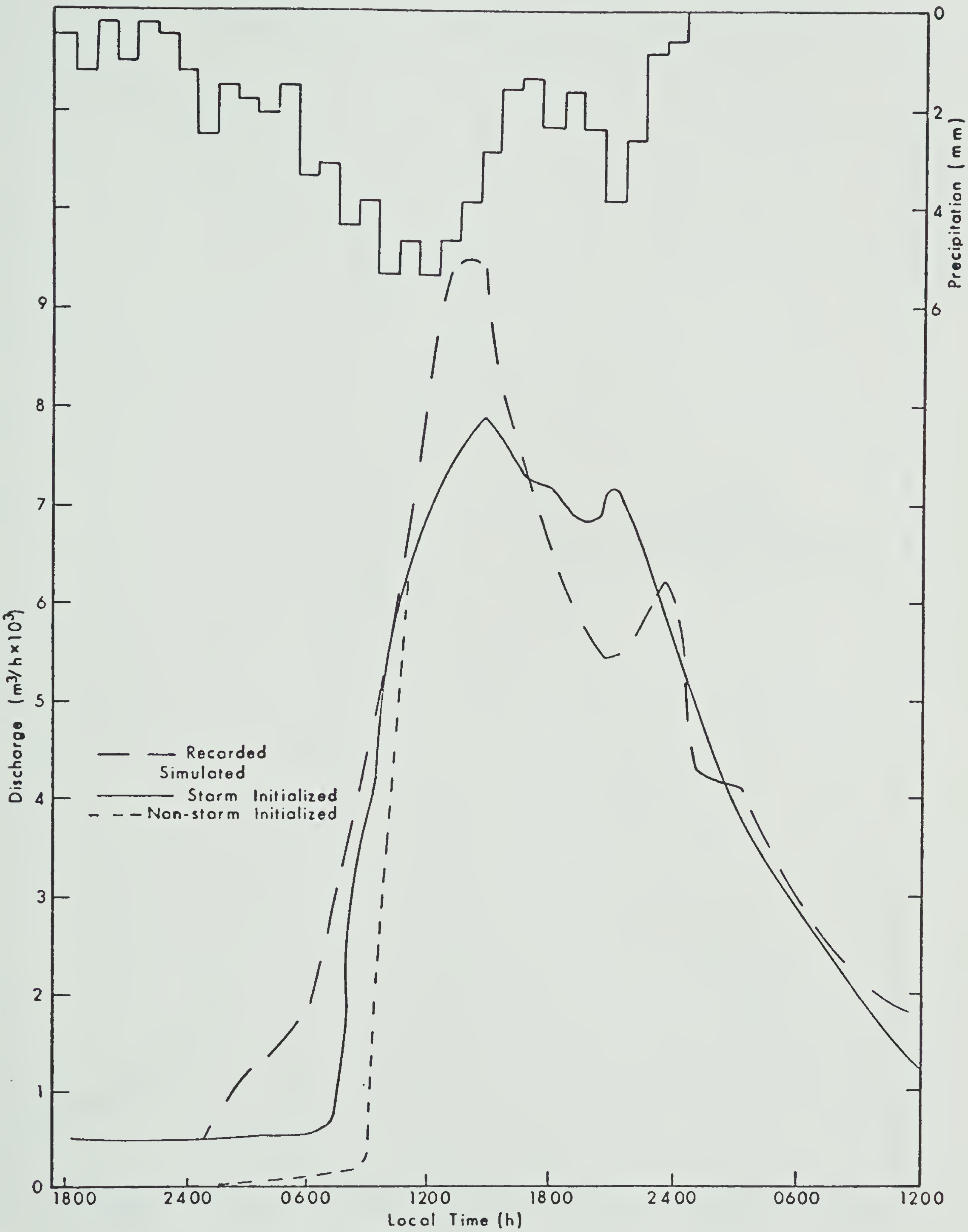


Figure 21c Initializing conditions and goodness of fit
Storm 8

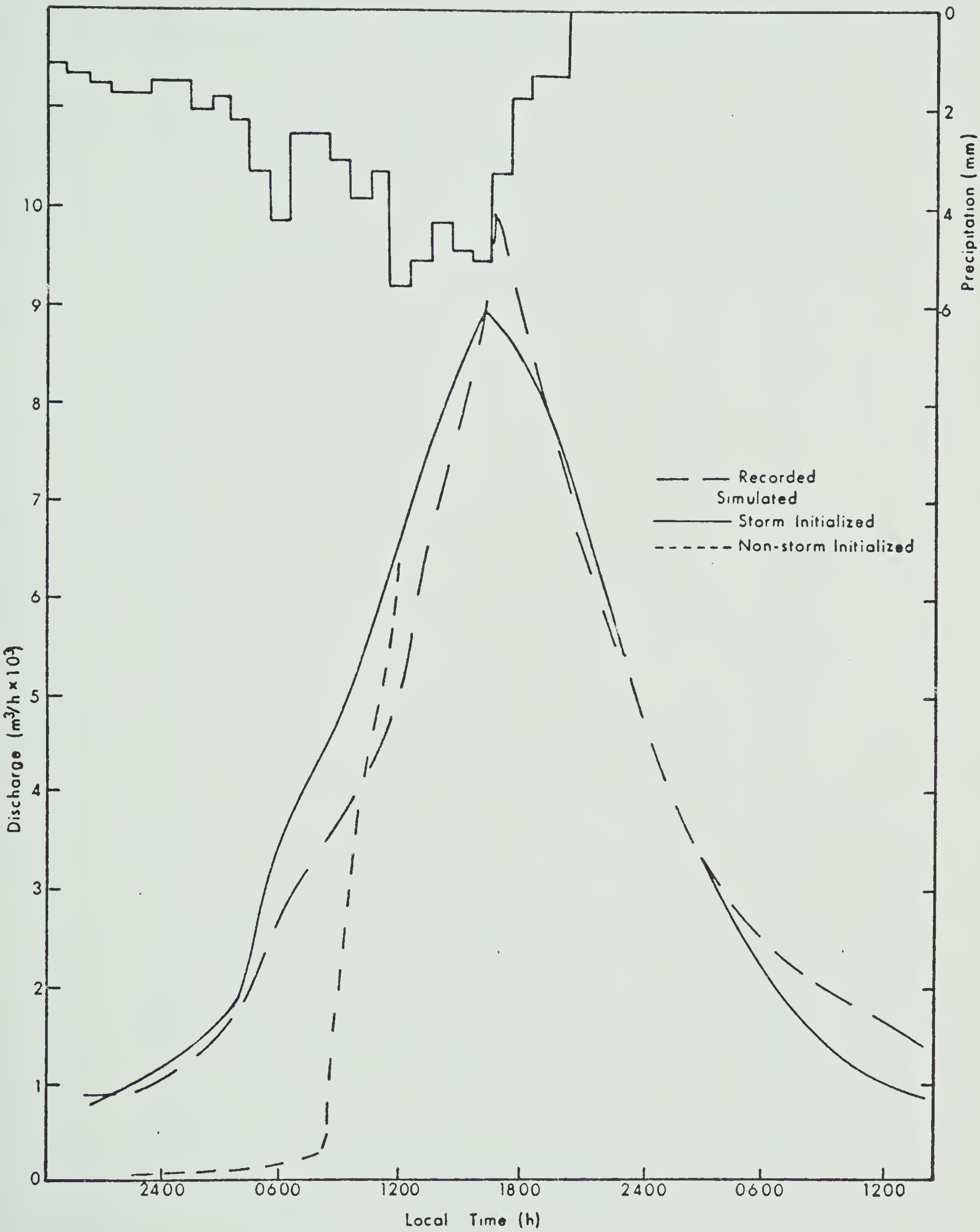


Figure 21d Initializing conditions and goodness of fit
Storm 11

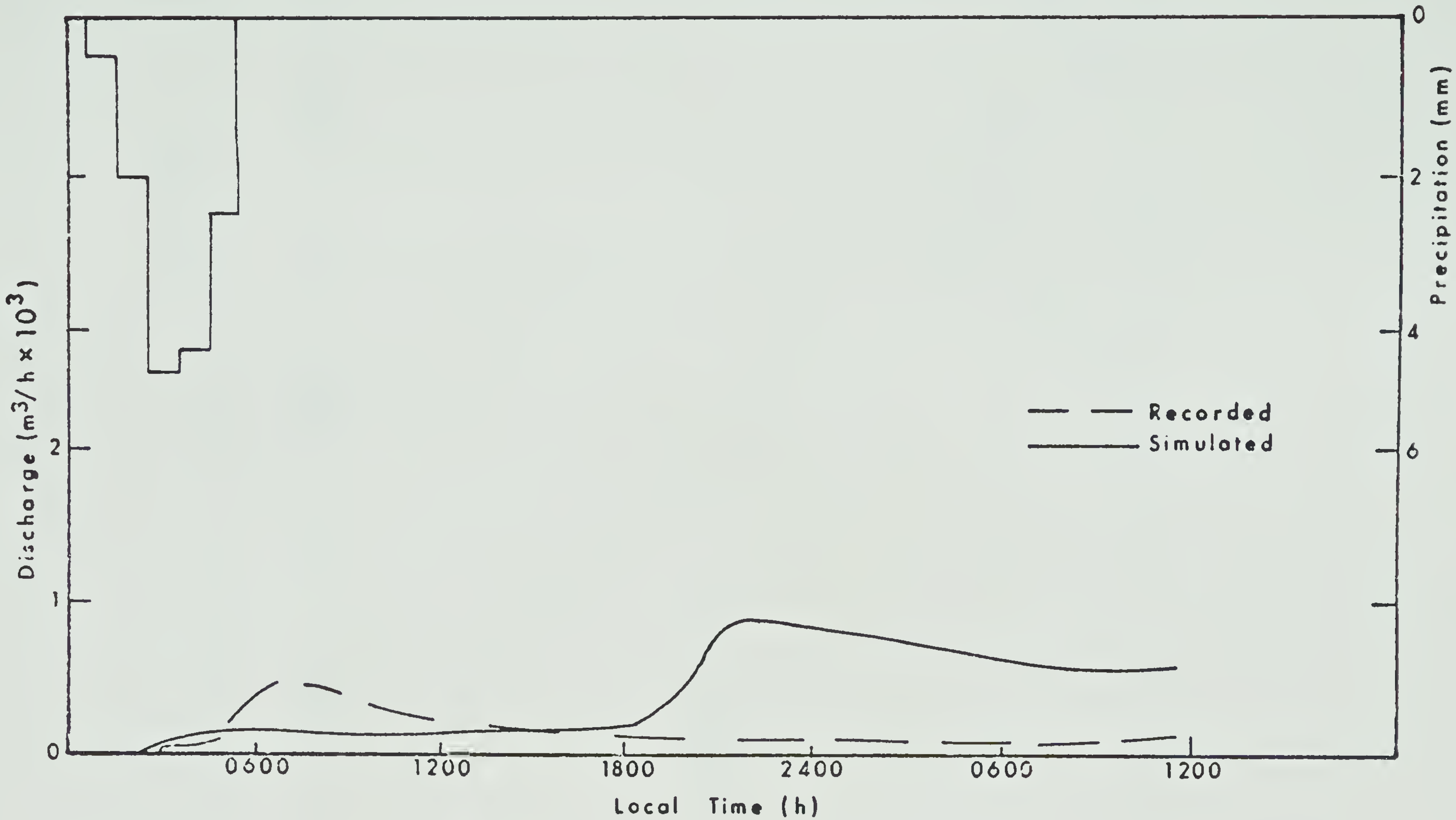


Figure 21e Initializing conditions and goodness of fit
Storm 35

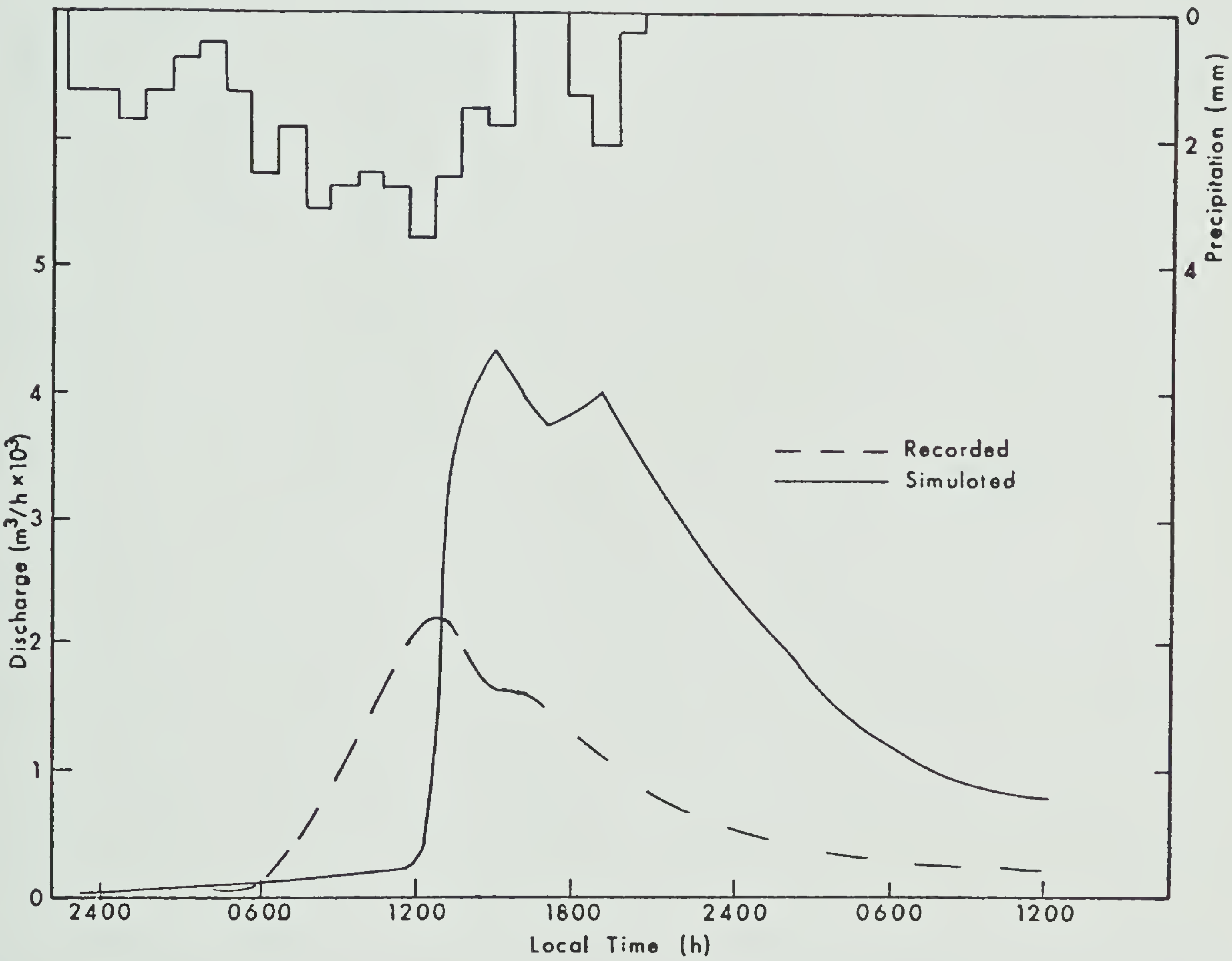


Figure 21f Initializing conditions and goodness of fit
Storm 36

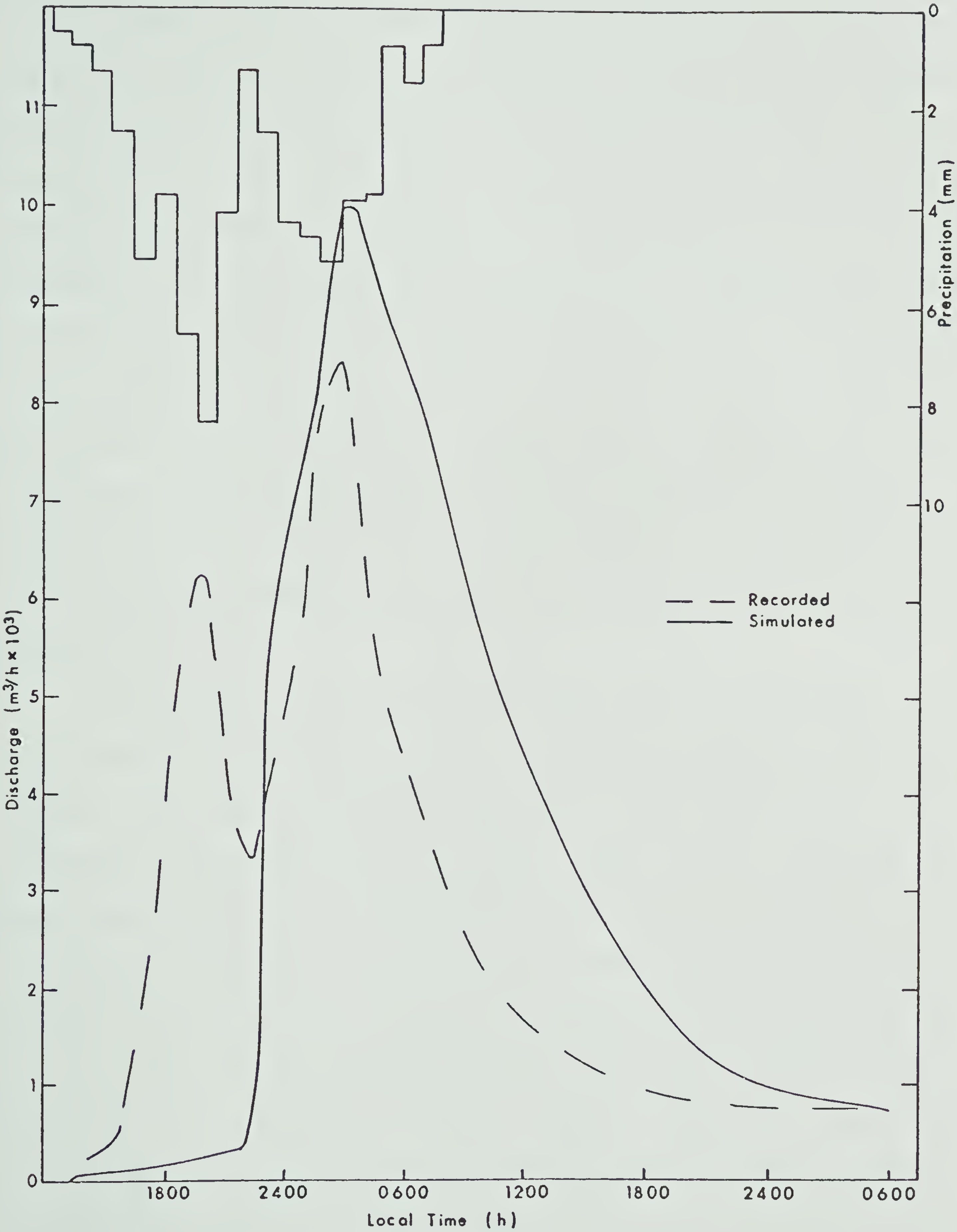


Figure 21g Initializing conditions and goodness of fit
Storm 37

limb of the hydrograph may be improved in certain instances through the use of a 'dummy' storm for initializing purposes.

Comparison of the recorded and simulated results using the 300 element grid indicates that agreement ranges from quite good for storm 4 to poor for storm 37. The simulated peaks for storms 2, 4, 36, and 37 indicate that the optimized conductivity coefficient might be too high. The simulations for storm 8 revealed that the model was responding to a second burst of rainfall but that it was responding too quickly (at a time when the simulated discharge was too high). A similar situation occurred in storm 36 but the recorded results did not show any response to this second burst of rainfall. This is in contrast to storm 37 which had the actual results showing a double peak but for which the simulated results did not respond to the first peak at all.

The results suggest that the model did a much better job of simulation for storms with wet antecedent conditions than for those with dry antecedent conditions. This result is to be expected since the model has been shown to be sensitive to antecedent moisture conditions. For storms 8 and 11, where the antecedent conditions were moist, a small change in moisture content would not be as critical as it would be for storms 35, 36 or 37 where the antecedent conditions were dry. A small 'error' in initializing moisture content would affect the simulated discharge hydrograph much more for these storms. The simulated results for storms 36 and 37, particularly on the rising limbs of the hydrographs, suggest that the initialized moisture contents

for both storms were possibly too low and, as a result, too much of the initial rainfall was being used to satisfy soil moisture resulting in a delayed rise in the simulated hydrograph relative to the recorded one. The simulations for these storms were initiated at their recorded baseflows, which were very low (see Table 2), suggesting that problems could be expected for the initializing of moisture contents for storms of low antecedent baseflow because of the sensitive response to moisture changes at the start of the storm.

8.2.2.8 Continuous Simulation

Until this point the capabilities of the model in regards to the simulation of specific storms have been discussed. This was done deliberately in an attempt to assess the physical 'correctness' of the model with respect to modelling of subsurface flow.

However, hydrologic models are more commonly used on a longer term basis, usually for an entire growing season. As an attempt to assess the capabilities of the model on a continuous basis, hourly simulations were made for the periods October 20 - November 7, 1974 and September 13 - 28, 1972. Pan evaporation data is not available for the watershed so daily potential evapotranspiration data was input as a constant value of 0.0625 mm/h. This value corresponds to an evapotranspiration value of 1.5 mm/day, a value obtained from Black _et_ _al_. (1973) for the months of October and November for the study area. Potential evapotranspiration was arbitrarily reduced to half the

input value if precipitation exceeded 2.5 mm/h and was reduced to zero if precipitation exceeded 4.5 mm/h. Actual evapotranspiration was linearly reduced from the potential value to zero as moisture content varied from field capacity to wilting point.

Comparison of the simulated discharges and the recorded instantaneous discharges for the period October 20 - November 7, 1974 is shown in Figure 22. The agreement is fairly good, with the exception of the double peaked storm 37, which occurred on November 5 - 6. Agreement between storm recessions is very good. A similar comparison for the period September 13 - 28, 1972 is shown in Figure 23 (note the change in scale between dates). Again the agreement between the simulated and recorded discharges for the recession period (September 22 - 28) is excellent. Agreement for the one major storm of the period which occurred on September 20 - 21, is poor with the simulated peaks being almost three times the recorded one. Considering the large amount of precipitation which had occurred until the time of peak discharge (78 mm), the recorded peak seems low, even for dry antecedent conditions. Furthermore, the recorded discharge hydrograph seems uncharacteristically flat. It may be possible that the recorded measurements during peak discharge were in error. In general the model demonstrated its ability to simulate discharge on a continuous basis quite adequately.

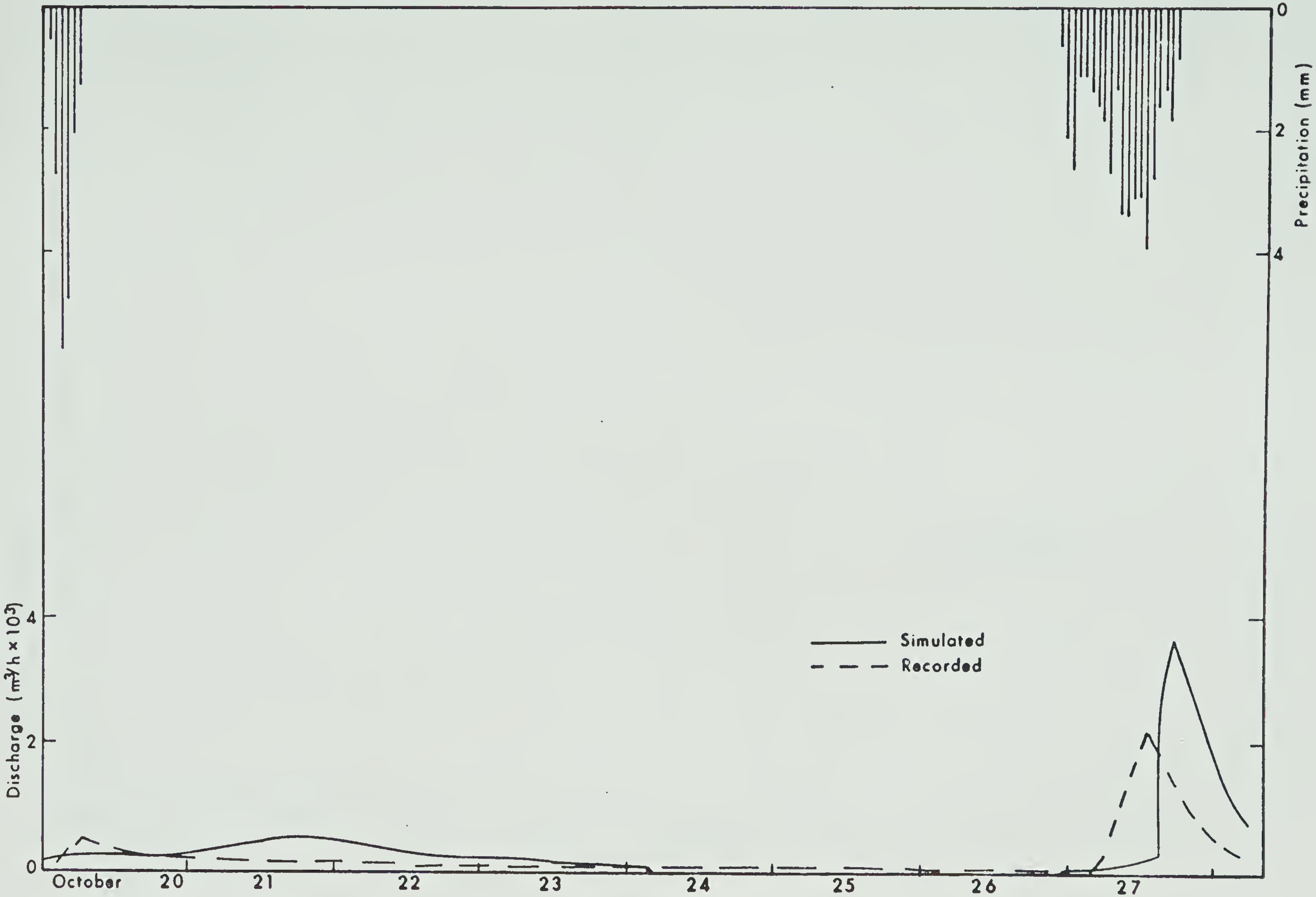


Figure 22a Continuous Simulations
October 20 - 27, 1974

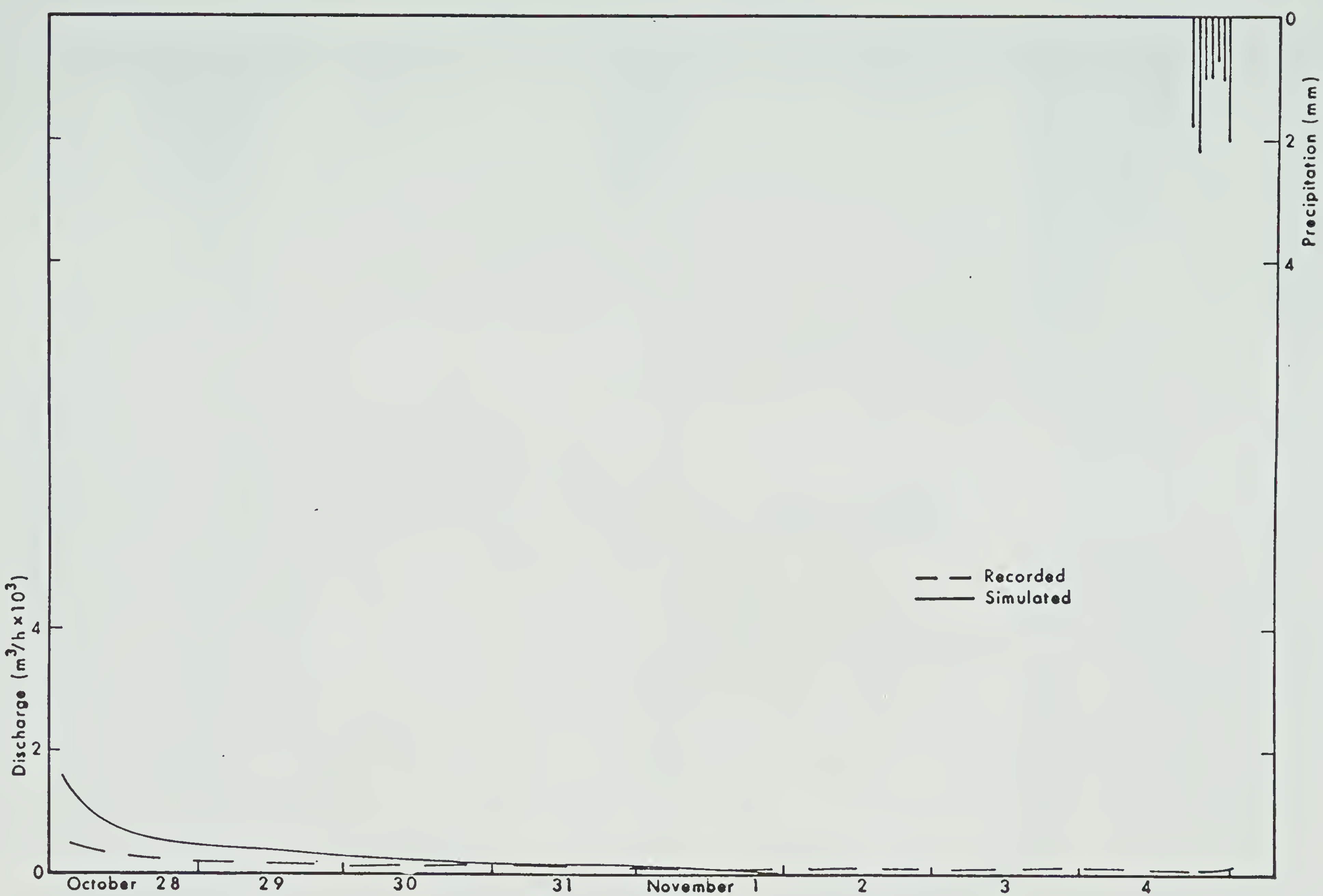


Figure 22b Continuous Simulations
October 18 - November 4, 1974

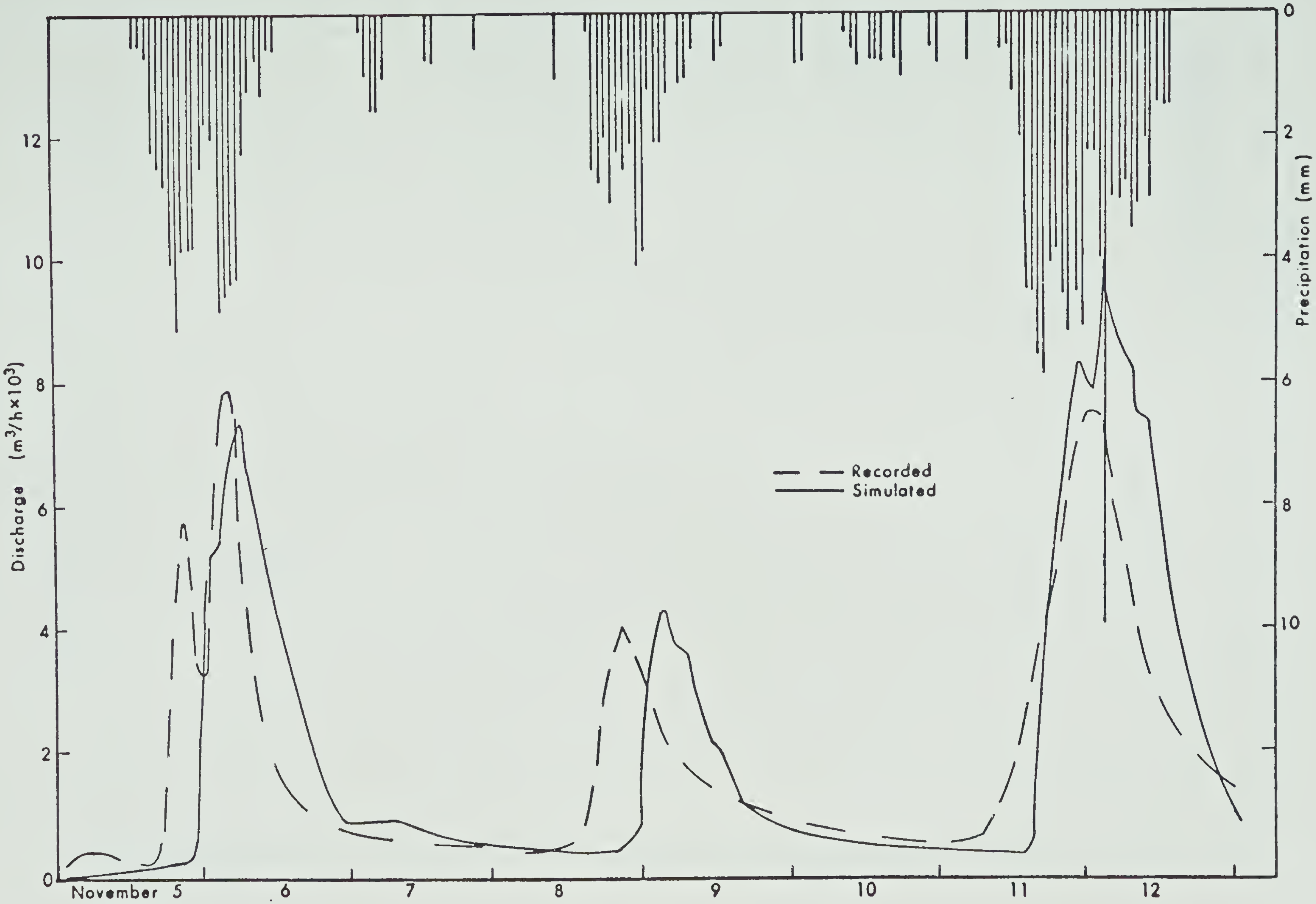


Figure 22c Continuous Simulations
November 5 - 12, 1974

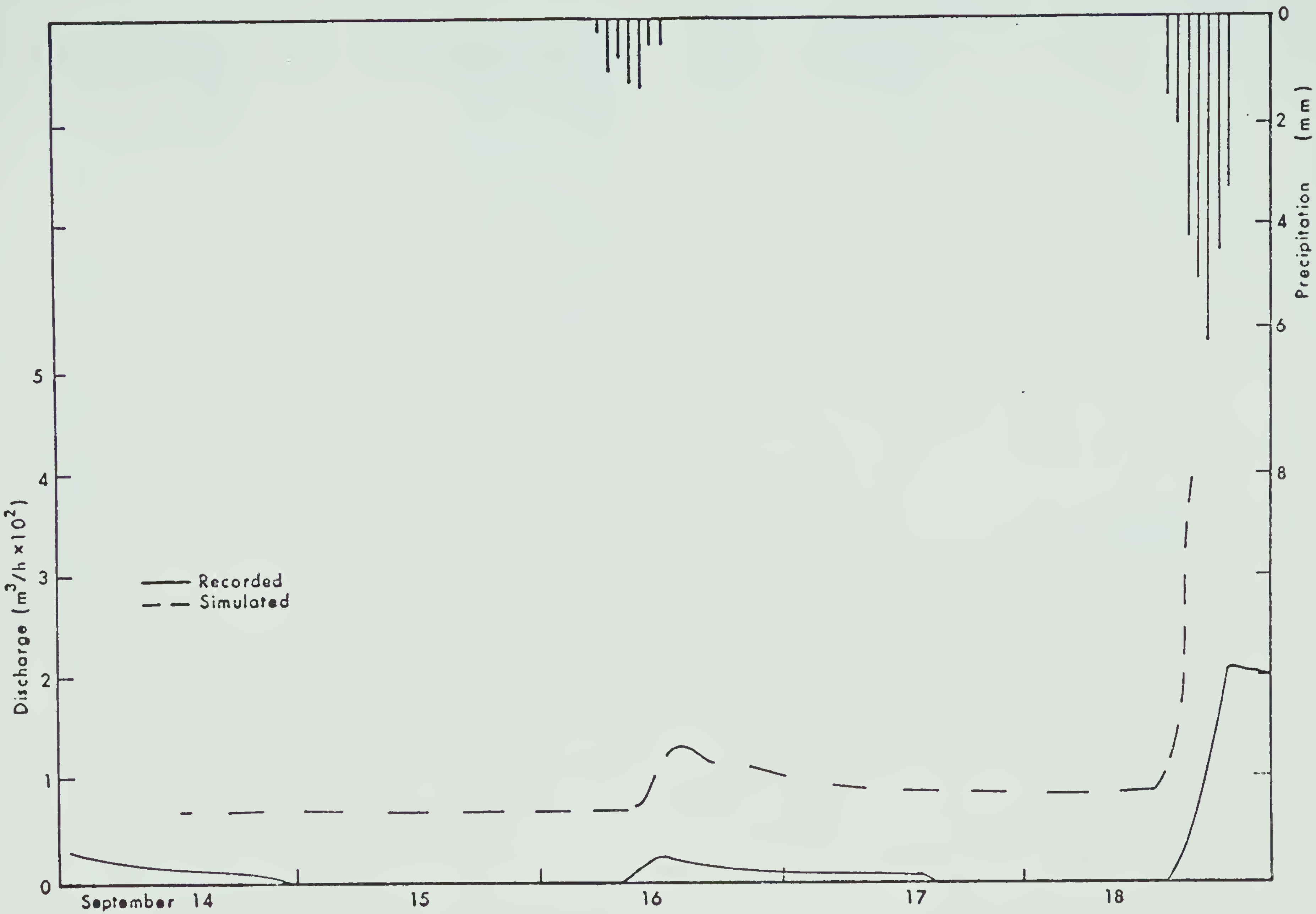


Figure 23a Continuous Simulations
September 14 - 18, 1972

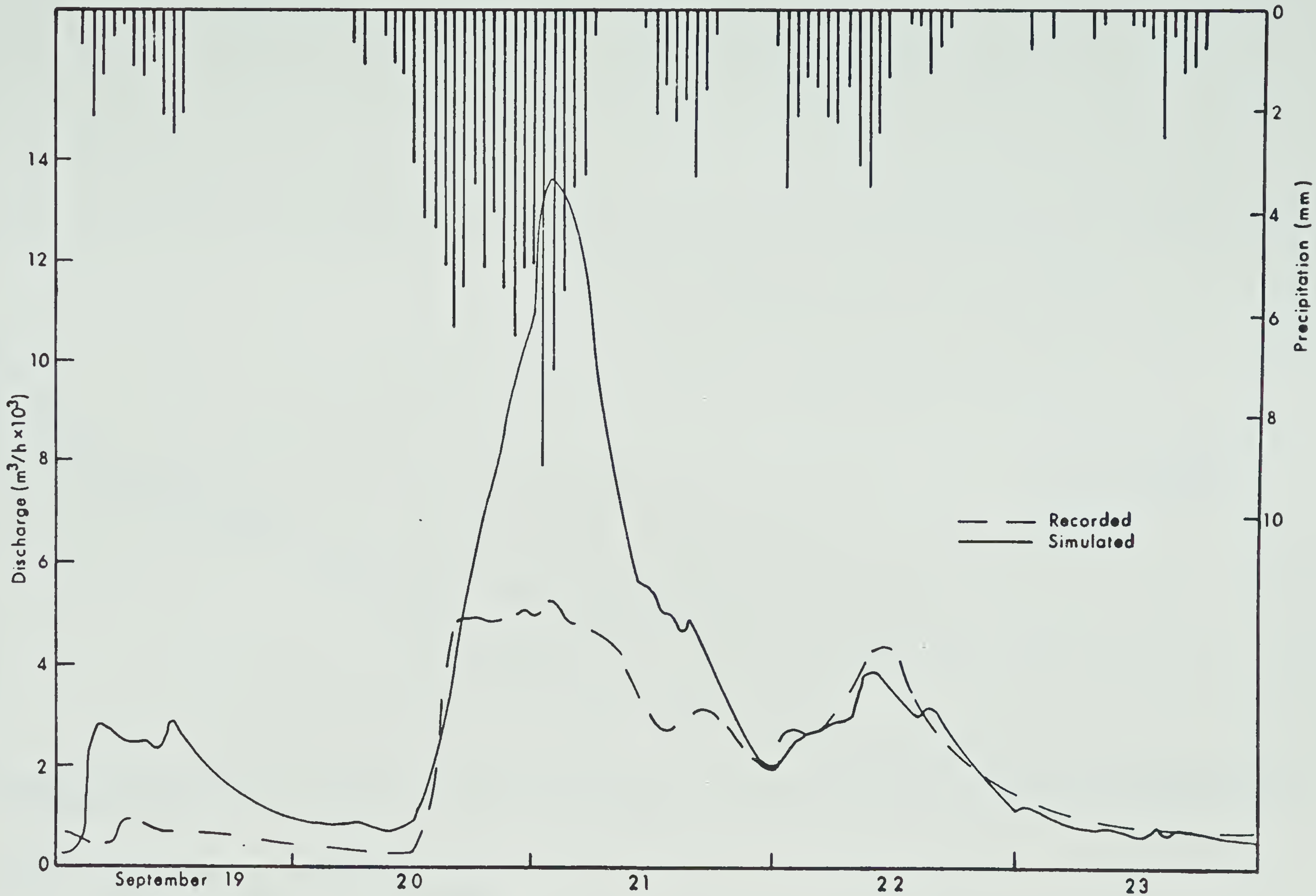


Figure 23b Continuous Simulations
September 19 - 24, 1972

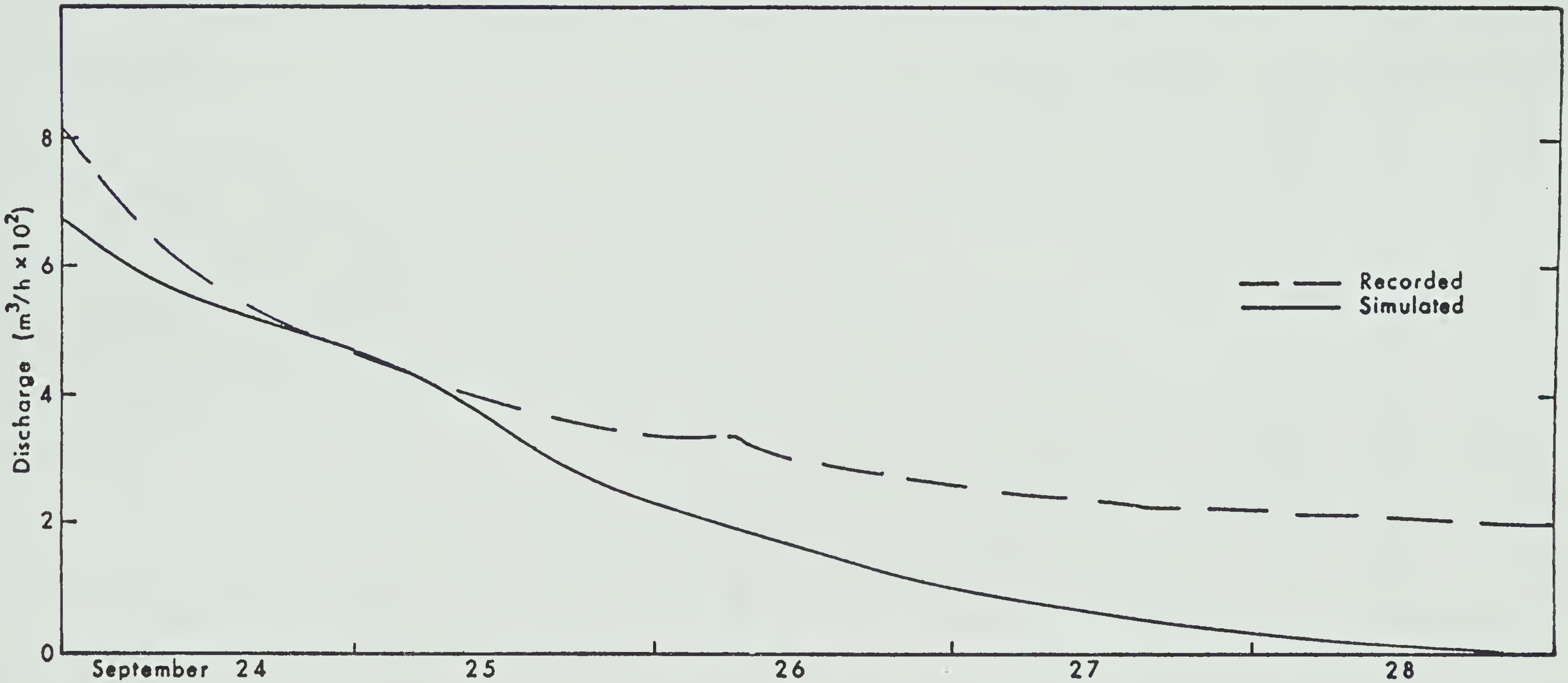


Figure 23c Continuous Simulations
September 24 - 28, 1972

CHAPTER 9

APPLICATIONS OF THE SLUICES MODEL

9.1 Application of the SLUICES Model To Watershed Management

The importance of watershed managers having a tool to predict the hydrologic response of a watershed due to watershed management practices has been previously discussed. The capabilities of the SLUICES model in this regard will now be examined.

9.1.1 Vegetation and Watershed Management

The role of watershed management with respect to water lies in improving the quantity and timing of water yields. The ultimate aim of watershed management in this regard is to maintain a more consistent streamflow throughout the year by reducing streamflow during periods of excessive rainfall and increasing it during periods of low rainfall. Vegetation is beneficial on watersheds because it maintains conditions in the soil that are favorable for infiltration and water storage and prevents erosion, but may be undesirable because it removes water from the soil thereby increasing the amount of rainfall required to wet the soil back up to the moisture content where it will begin to discharge water. Therefore, the usefulness of vegetation is generally determined by the relative demand for

water, weighed against the hazards of floods and erosion.

The first well-documented, deliberate experiment of changing land use in order to quantify the effects on streamflow was undertaken at Wagon Wheel Gap in Colorado during the 1910 - 1920's. Hibbert (1967) reported results for thirty-nine studies of the effect of altering forest cover on water yield carried out since that time. Collectively these results reveal that forest reduction increases water yield and that reforestation decreases yield. In humid regions, streamflow response is proportional to the reduction in forest cover. As the forest regrows following treatment, increases in streamflow decline. As well, the seasonal distribution of streamflow response to treatment is variable depending upon climate, soils, topography, etc.

If the soil is permeated by roots of transpiring plants, water is removed until the wilting point is approached. The soil must then be wetted to field capacity before it will begin to discharge subsurface flow. If a soil is at field capacity when precipitation occurs, whenever water enters, it will release an equivalent amount into the subsurface flow system. Since vegetation can remove practically all the readily available water to a depth of the root system, precipitation is required to replenish this water before substantial flow will occur. In the absence of vegetation, a smaller loss of water would occur because in most soils the loss by evaporation occurs largely from the surface soil horizon; hence, a smaller amount of rainfall would be required to recharge the soil to field capacity. If less

water is depleted from a clearcut area than from a nonlogged one, then less water is required to return the soil mass to field capacity and consequently a greater portion of precipitation becomes available for streamflow. The amount that evapotranspiration can be reduced and streamflow increased by watershed management depends upon the management practice employed and the amount of water available for evapotranspiration. Most investigators report that well-stocked forest vegetation appears to use water at about the same rate regardless of species and thus the kind of species plays a lesser role. The hydrologic response to forest treatment depends on the physical characteristics of the watershed, including the soils and the post-treatment recovery of the treated area. Changes in vegetative density by modifying the area of the transpiring surface affect evapotranspiration rates from forest stands. Reducing stand density reduces evapotranspiration and the greater the density reduction, the greater the evapotranspiration reduction (Bethlahmy, 1962; Douglass, 1960; McClurkin, 1961; and Zahner, 1958).

Roots of fully stocked stands are approximately evenly distributed horizontally. When trees or groups of trees are removed, the uniform pattern of rooting is interrupted, roots are concentrated near trees and few roots occur in openings. Continuing season soil moisture accretion will then occur in the thinned stand.

The plant, soil, and atmosphere are all integral parts of a single dynamic system which transfers water from the soil to the

atmosphere. Trimble et al. (1963) describe three possible combinations of these factors. Where growing season rainfall is sufficient in amount and time to supply the evapotranspiration need without much soil drying, heavy cutting can result in a large increase in water yield (Case I). If the growing season rainfall is low, evapotranspiration in the uncut stand will be limited by a drier soil, and the cutting of the trees will produce little increase in water yield since the limited rainfall can do little more than satisfy the evaporation loss from the soil (Case II). Under an intermediate condition of rainfall, after cutting, evaporation uses only a part of the soil moisture that would be evapotranspired under a full cover. The resultant higher level of soil moisture at the end of the growing season will result in an increase in water yield because less rainfall from subsequent storms will be needed to satisfy the soil moisture defecit (Case III).

Thus, maximum gains from forest cutting can be expected where year-round conditions of full potential evapotranspiration prevail, i.e. where precipitation results in maximum moisture stored. Where water supply cannot meet the full potential evapotranspiration requirement, there will be decreased opportunities for evapotranspiration savings. But clear-cutting a forest does not save all the water which would otherwise be transpired. Evaporation from bare soil and litter will still continue, but at a fraction of the amount evapotranspired under a full cover.

Trimble et al. (1963) also made note of the importance of

soil depth, since the soil represents a reservoir to be filled before rainfall can contribute to water yield. Within limits, under a rainfall situation as described in Case III, the deeper the soil, the greater the possibility of increasing streamflow. If the soil is wet, as in the Case I, differences in soil depth will have little or no effect due to cutting since the soil is already near field capacity.

Under certain conditions of rainfall, the depth of the soil may determine whether cutting produces any increase of streamflow at all. With a shallow soil (and therefore low soil moisture storage capacity), evaporation potential might not be satisfied even after cutting and there would be no increase in water yield. With the same rainfall but deeper soil (therefore more stored soil moisture), the evapotranspiration demand might be satisfied and Case III would exist. Soil profile depth also influences the length of time that increases in water yield due to cutting will persist. The deeper the soil, the longer it takes roots of new growth or the expanding root systems of the remaining plants to occupy the entire soil mantle. When the soil profile is reoccupied, transpiration reaches maximum levels again and increases in streamflow disappear.

Commercial clear-cutting of entire watersheds is rare. Usually only certain portions of the watershed are deforested at any one time, resulting in a checkerboard pattern on the watershed. The size of the clearcut area varies, depending on the treatment. Rational management of forested watersheds for the enhancement of water yield requires that the hydrologic

consequences of any given treatment be predictable, and modelling has been shown to be a useful tool in this regard. Partly distributed models based on areal averaging of alternate forested and clearcut areas must inherently assume that the relative location of the clearcut blocks within the watershed does not matter. Based on the earlier discussion on soil properties, the assumption seems a poor one. A distributed model is not limited by such an assumption. The SLUICES model is ideally suited for the investigation of the effects of partial clearcutting on soil moisture and streamflow because it is distributed in nature and has been shown to be sensitive to moisture content changes.

The measurement of evapotranspiration from a forest is difficult and thus this parameter is most often estimated indirectly from the difference between precipitation and streamflow during selected time periods, estimated from a water budget developed from periodic measurements of soil water and precipitation or estimated from aerodynamic or energy budgets.

The amount of water evapotranspired from mature forests also depends on the amount of moisture available in the soil, which in turn is largely a function of soil texture and soil profile depth.

Estimation of evaporation from clear-cut areas is very difficult. The nature of the forest treatment in the removal of trees can play a very important role in determining the rate of evaporation in the clear-cut area. Anderson et al. (1976) presented a data compilation on increases in water yield

following forest cutting for several geographic locations, forest types and types of cutting. Data from Western Oregon (similar general locality as Jamieson Creek) show that for a 96 ha watershed, a first year increase in water yield of 462 mm was experienced when the watershed was clearcut. Since mean precipitation was 2286 mm and streamflow was 1448 mm, post-harvest evaporation was 45% of the pre-harvest evapotranspiration level. When a 101 ha watershed in the same area was 30% clearcut, a first year water yield increase of 150 mm was experienced. Thus evaporation on the clearcut area was 40% of that which remained forested. Based on these figures, an initial value for clear-cut evaporation that is 45% of the forested evapotranspiration will be used.

9.1.1.1 Effects of Watershed Management on Soil Moisture

To test the ability of the SLUICES model to evaluate the effects of clear-cutting on soil moisture and streamflow, simulation runs were made using meteorological data from Jamieson Creek watershed for the period October - November, 1974. Daily potential evapotranspiration data was input at a constant value of 0.0625 mm/h as discussed previously in Section 8.2.2.8. Model simulation was carried out for 720 hours and the time period included storms 35, 36 and 37. A 'dummy' storm was used to initialize the moisture conditions on the watershed. Two simulations were run: one with the watershed trees evapotranspiring at a rate discussed above and the other with all the trees removed and the evapotranspiration rate set at a

value of 45% of the evapotranspiration rate that would exist if trees were present. Moisture contents for all elements were examined 358 hours into the actual simulation (53 hours after the end of storm 36). The moisture content of each element for the 'treed' simulation was then compared to the moisture content of the same element under the 'no trees' simulation. Results of the comparison are shown in Figure 24. Note that each data point may actually represent several elements. The 45^{o} line indicates cases of no change in moisture content due to clearcutting. The moisture contents for all elements under the 'no trees' simulation can be seen to be consistently higher than those under the 'treed' simulation. Although the differences in moisture content are very small, the model simulations are encouraging, considering the rather short time span of simulation. Subsequent increases in streamflow could be expected.

9.1.1.2 Effects of Watershed Management on Streamflow

To investigate whether in fact the SLUICES model was capable of showing expected increases in streamflow due to clearcutting, seven different simulations were run with a range in percent area clearcut from 0 to 100% with intermediate values at 73.7, 48.3, 24.3, 11.0 and 4.3% with complete clearcutting and no clearcutting options also simulated. Precipitation data were synthesized from that for Jamieson Creek with the total precipitation for the 30 days of simulation considered being 308 mm. This value would correspond to an average monthly precipitation total lying between the values for September and

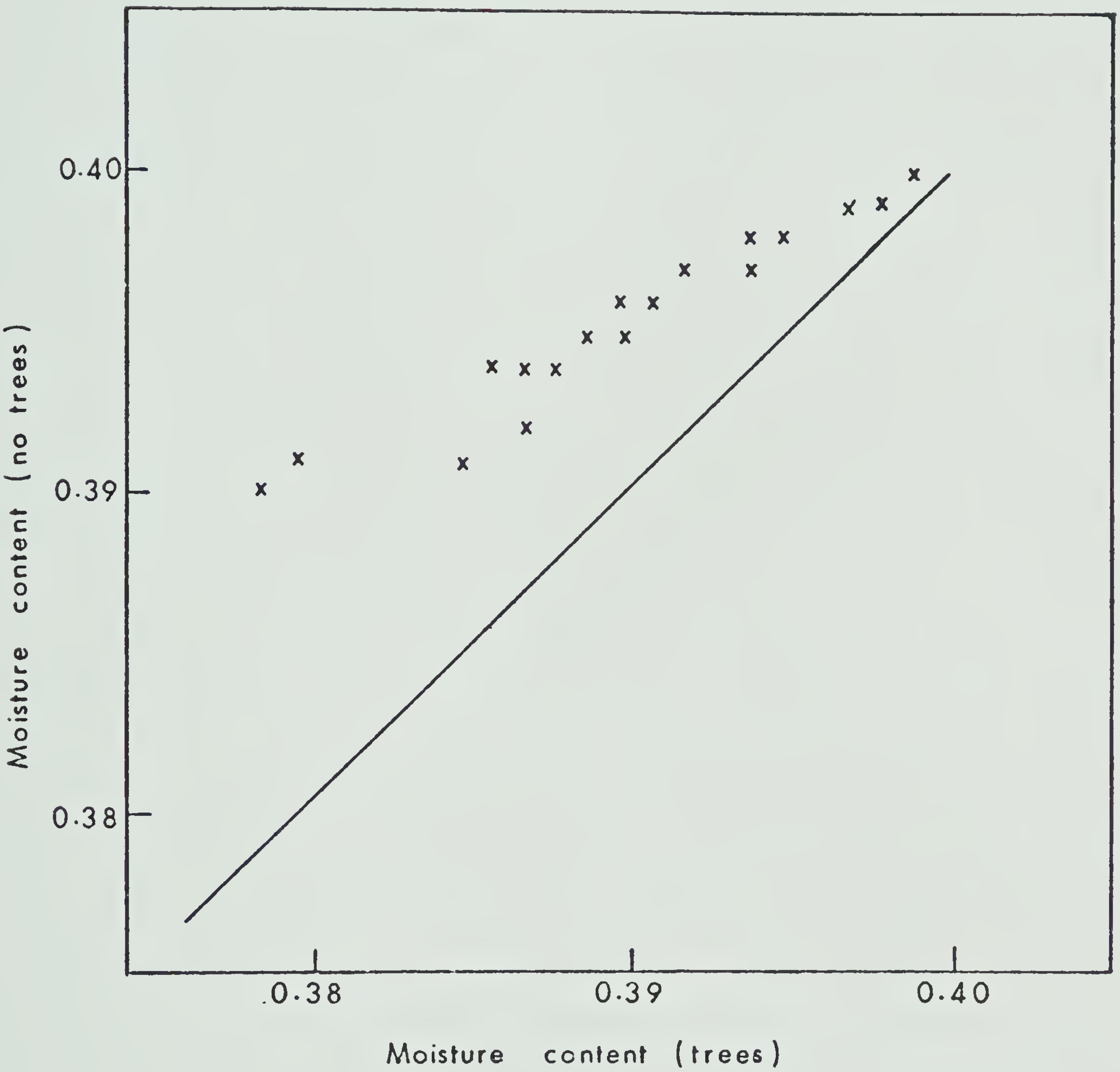


Figure 24 Soil moisture status: Trees versus no trees

October, as reported for Seymour Falls (see Figure 13). Evapotranspiration was incorporated in the simulations in the same manner as discussed previously, with the same reduction for clearcutting being used. Simulations were run for 960 hours but the first 50 hours were not included in the analysis to allow for initializing of moisture conditions (storm 11 was used). Beginning with the 51st hour (discharge about 610 m^3/h), the simulations were then broken into thirty 24 hour periods. To examine the effect of clearcutting on discharge, hourly discharges were summed to give daily values and then these were summed for a thirty day period. The 30 day total volume for the simulation in question was compared with the corresponding total for the completely treed watershed simulation and recorded as a percent increase. Results are shown in Figure 25. The points for upstream areas are for clearcut areas at the uppermost points of the watershed and working systematically downslope by columns. The results suggest that for the given thirty day period, a 7.7% increase in overall discharge volume would result if, under the given meteorological conditions, the watershed were clearcut. For clearcut areas smaller than the total basin area, the percent increase is reduced proportionately. The results were very consistent in that every treatment, on every day, gave higher daily discharge volumes when compared to the completely treed simulation. The per cent increase in daily discharge, as compared to the completely treed simulation, varied from 1.3% to 184%, with an average increase in daily discharge of 20.1%.

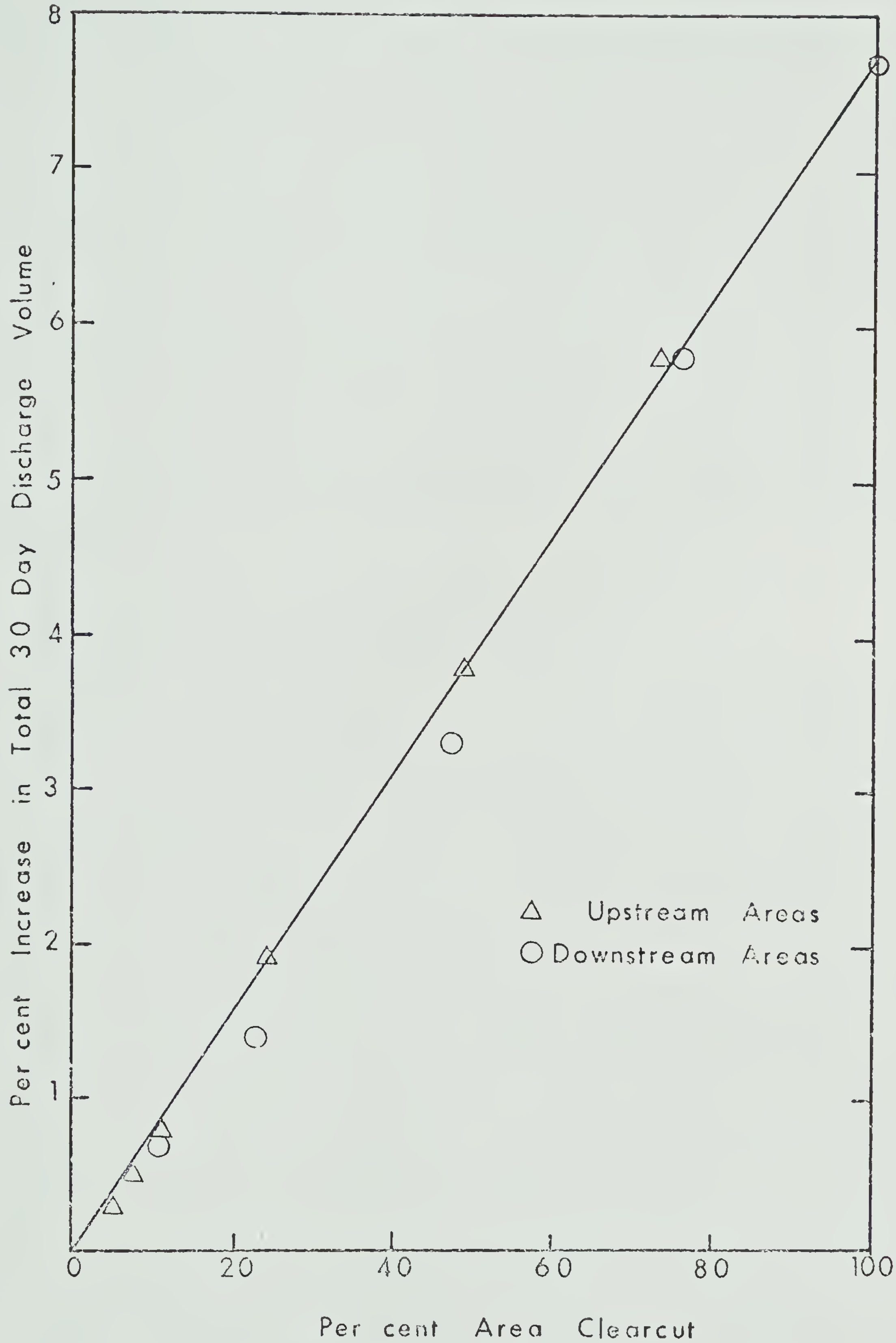


Figure 25 Effect of clearcut area on total discharge

In an effort to assess the effect of clearcut location on streamflow increase, areas of 75.7, 47.3, 23.0 and 10.3% were clearcut but the areas now originated at the outflow point. The results are also shown in Figure 25. If the slight difference in areas were to be considered, the overall effect of location would be negligible. This may be largely due to the small size of the watershed. Examination of the daily discharge totals showed that the downstream clearcut areas showed slightly quicker response to rainfall than did an equivalent upstream clearcut area but the trends were not always clear since the daily ratios of comparable discharges, upstream and downstream, varied on both sides of equivalence.

9.1.2 The Importance of Evapotranspiration Level

To assess the sensitivity of the results to the level of evapotranspiration, clearcutting simulations were repeated using a daily value of 5 mm/day, up from 1.5 mm/day. This value was taken to represent an upper level of evapotranspiration for Jamieson Creek and would be characteristic of this watershed during the months of June, July and August (Black <u>et</u> <u>al</u>., 1973). Results of the simulation indicated that, for a daily evapotranspiration rate of 5 mm/day, clearcutting of the entire watershed, for a 30 day period of simulation using the same meteorological conditions as discussed earlier, resulted in an increase in total volume of discharge of 36%, as compared to 7.7% with an evapotranspiration rate of 1.5 mm/h. Also of interest is the fact that the total 30-day volume of discharge for an evapotranspiration rate of 5 mm/day for an entirely treed

watershed was only 69% of that for an evapotranspiration rate of 1.5 mm/day for the treed watershed. The total volume of discharge for the clearcut watershed for an evapotranspiration rate of 5 mm/day was 87% of that for similar conditions except for an evapotranspiration rate of 1.5 mm/day. These results indicate the greatest percentage increase in volume of discharge due to clearcutting would be felt on watersheds where the evapotranspiration rate is high. This also suggests that, for Jamieson Creek, for comparable storms, a greater increase in discharge would occur during the summer months than in the fall.

Another evapotranspiration factor which must be considered is the proportion of completely treed watershed evapotranspiration that occurs after clearcutting. Until this point a factor of 0.45 (45% of completely treed evapotranspiration occurring in clearcut areas) has been used. Factors of 0.25 and 0.65 were also examined with simulations for complete clearcutting being made using a daily evapotranspiration rate of 5 mm/day. The percentage increase in total volume of discharge for a 30-day period for factors of 0.25, 0.45 and 0.65 were 49%, 36%, and 23% respectively, indicating that the greater the reduction in evapotranspiration after clearcutting, the greater will be the increase in volume of discharge experienced. This conclusion is based on the assumption that clearcutting does not disturb the soil's water-holding, transmitting or infiltration properties in any way. These results reflect the often reported finding that the per cent increases in yield cannot be sustained at its first year level without some form of growth retarding measure. The

values for evapotranspiration reduction of 0.25, 0.45 and 0.65 reflect these effects of the stages of regrowth on the hydrologic response and show that the amount of evapotranspiration reduction can significantly affect the hydrologic response of the watershed.

9.1.3 Effects of Watershed Management on Storm Hydrographs

To gain information about the magnitude of storm hydrograph most affected by clearcutting, hydrographs for the total 37 day period of simulation were plotted and are shown in Figures 26 and 27. Storm peak increases due to clearcutting are more evident for an evapotranspiration rate of 5 mm/day than for a rate of 1.5 mm/day. Two storms were examined more closely. The first storm occurred on the eighteenth day of simulation and the other on the thirty-seventh (see Figures 26 and 27). Total storm precipitation for the two storms was 29 and 96 mm respectively. The storm with lower precipitation had antecedent flows approximately one third those of the larger storm. Hydrographs for the two storms for the two management options of entirely treed and clearcut for two evapotranspiration rates of 1.5 and 5 mm/day are shown in Figures 28 and 29. Note the change of time scale with respect to the previous two figures. The most noticeable effects on the discharge hydrographs due to clearcutting are evident for the smaller storm with the higher evapotranspiration rate showing a more significant change in discharge. The hydrograph for the larger storm, regardless of the option considered, varied little, with only a slight effect

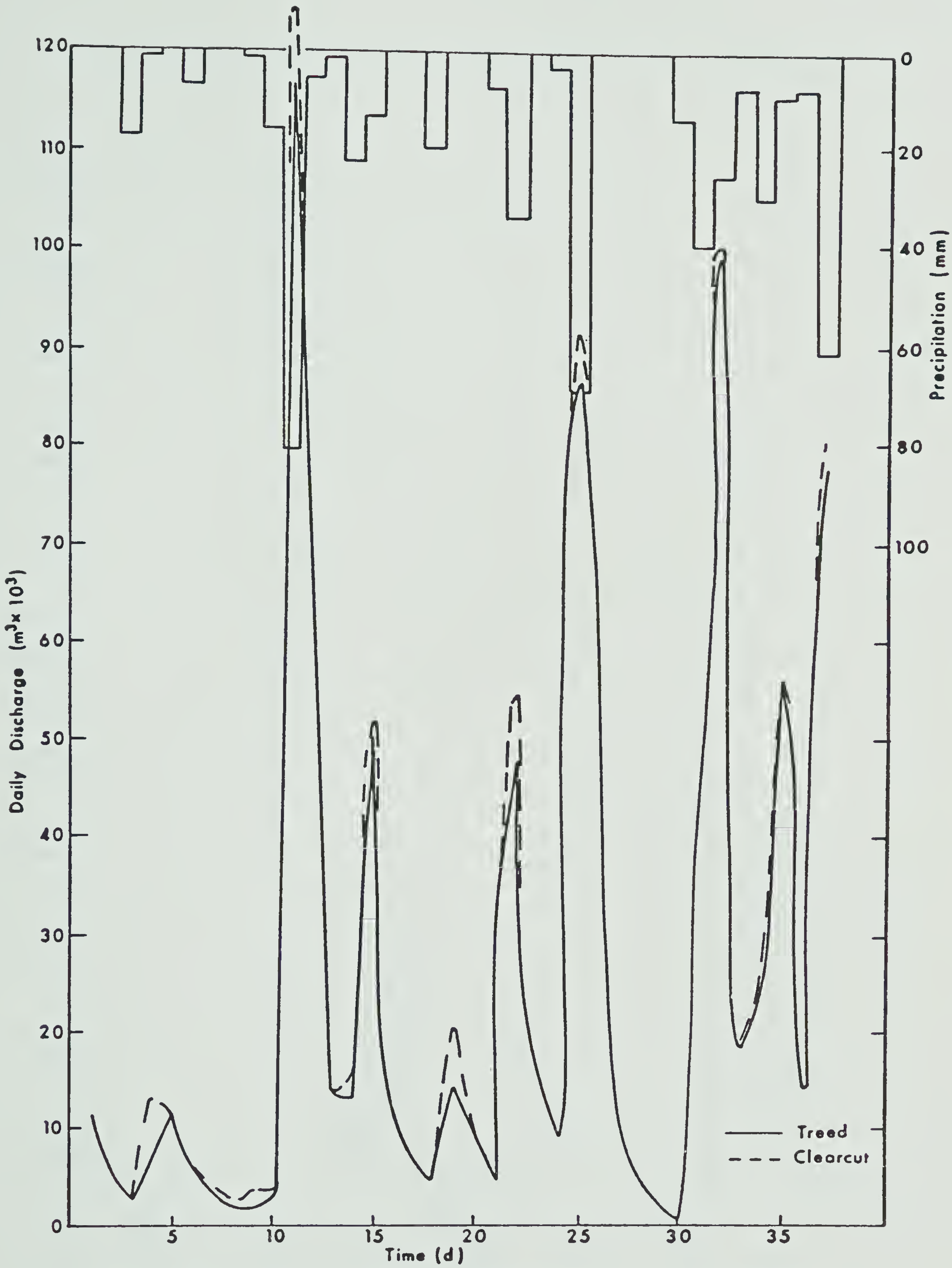


Figure 26 Simulated hydrographs for evapotranspiration rates of 1.5 mm/d

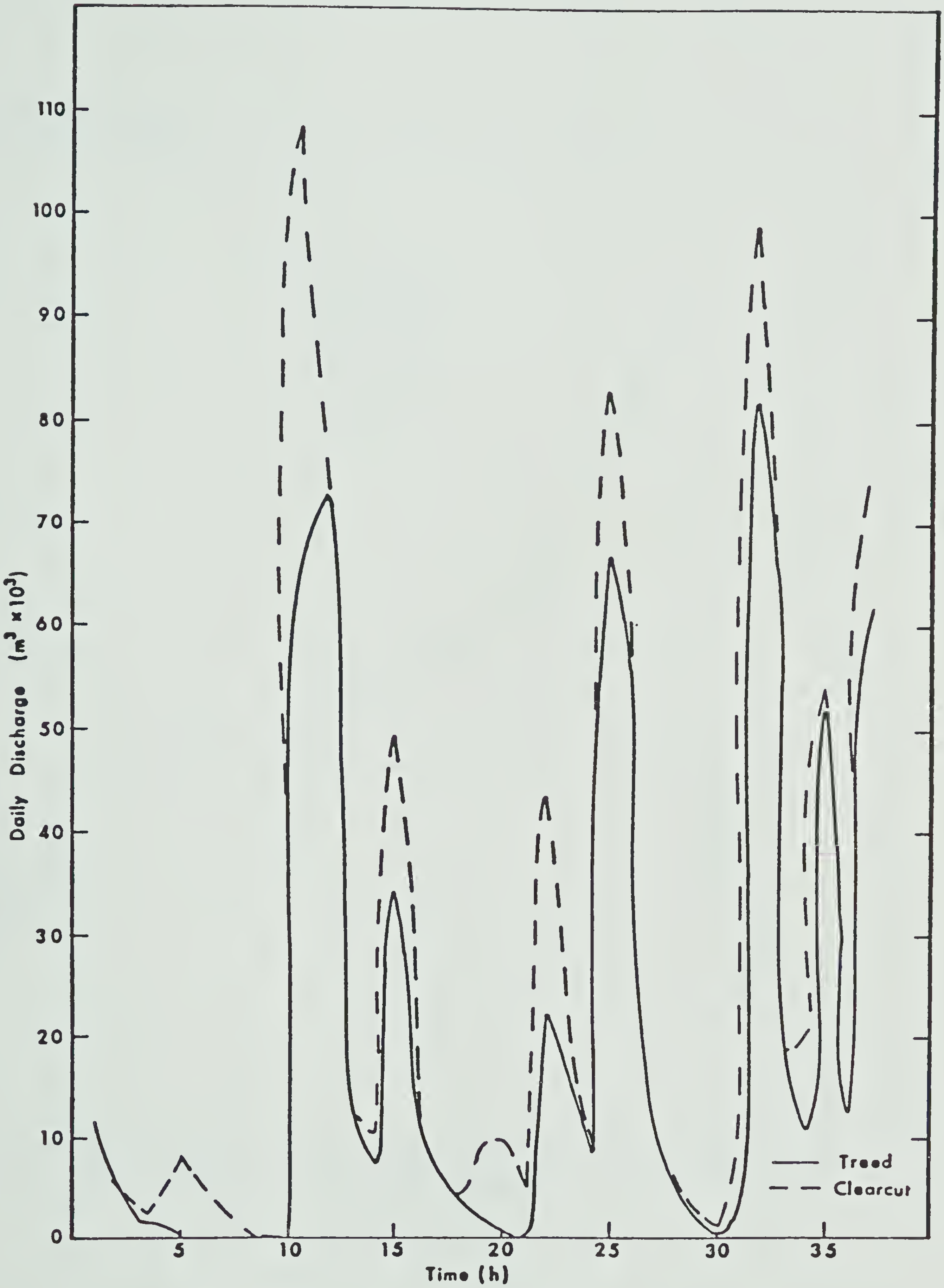


Figure 27 Simulated hydrographs for evapotranspiration rates of 5.0 mm/d

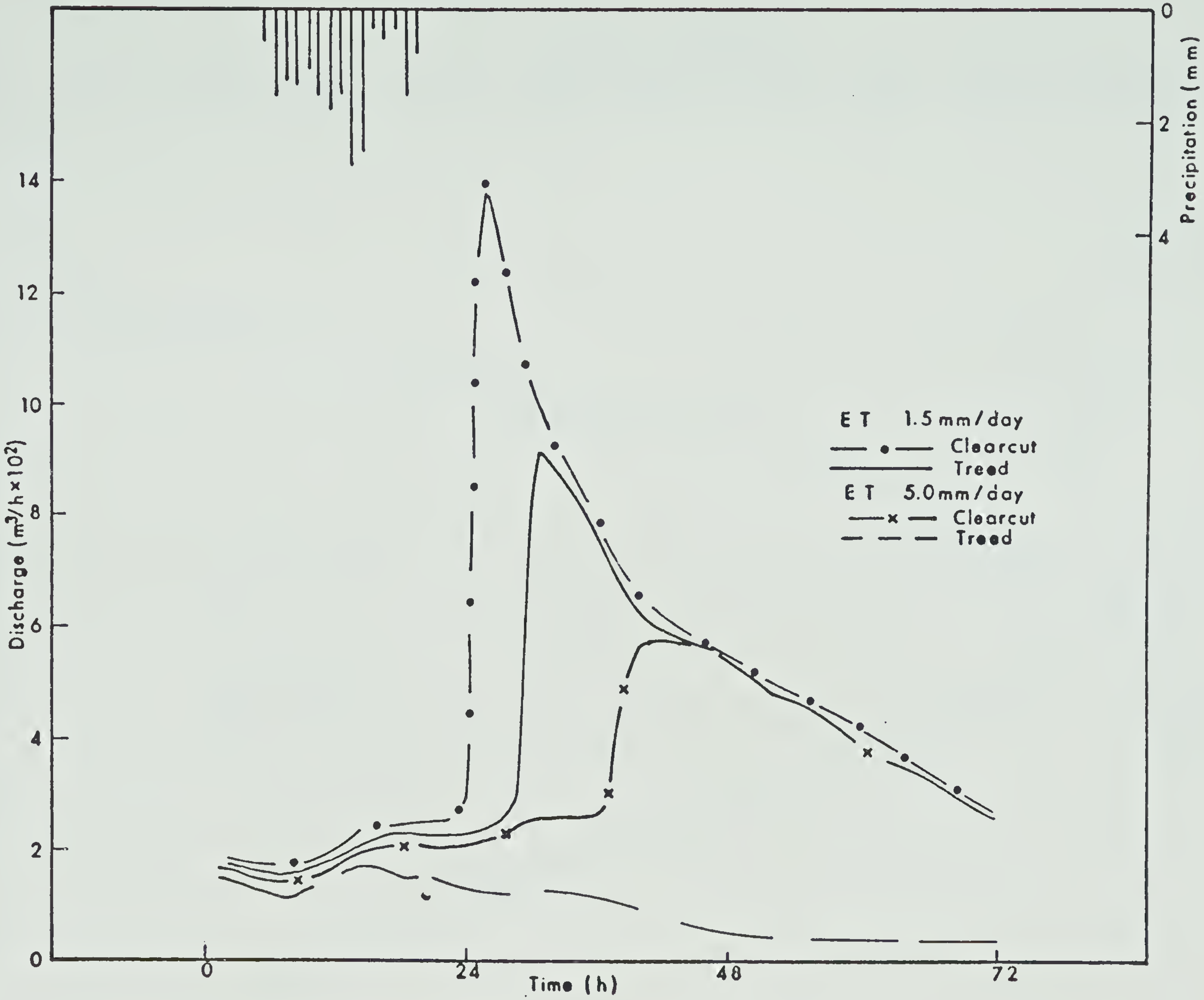


Figure 28 Discharge hydrograph for a storm of 29 mm

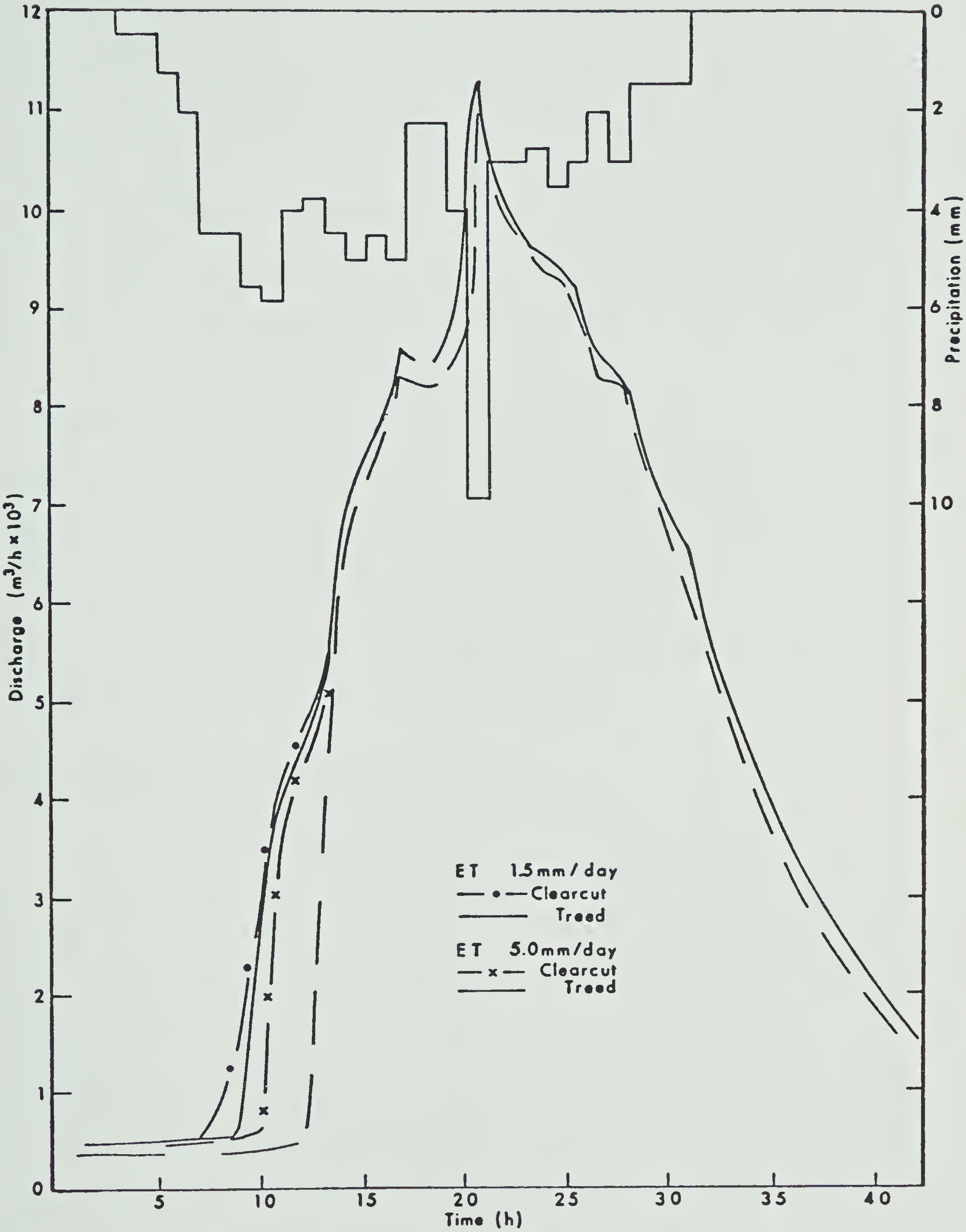


Figure 29 Discharge hydrograph for a storm of 96 mm

shown on the rising limb. The results suggest that the effects of clearcutting will be felt most for the small to intermediate size storms, while the effect on discharge from large storms would be minor. These simulated results correspond to what one would expect to happen in nature. Of prime importance in this regard is the proportion of precipitation that must be used to satisfy soil moisture before storm discharge can begin. For the small to intermediate size storms, this proportion can be significant, and a management practice, such as clearcutting, can increase discharge. However, for large amounts of precipitation, the proportion of it used to satisfy soil moisture defeicts is small, and thus, even though clearcutting increases the soil moisture, the decreased proportion of precipitation required to satisfy soil moisture is not significant enough to markedly affect the hydrograph or the total volume of discharge.

9.1.4 Comparison of Simulated Results with Literature Reported Results

Numerous studies of the effects of forest clearcutting on streamflow have been conducted in North America since the initial study was conducted by Bates and Henry in the 1920's. Hibbert (1967) reported the results of thirty-nine studies aimed specifically at evaluating the effects of forest cover manipulation on water yield. His generalization from the studies was that reduction of forest cover increases water yield but that the response to treatment was highly variable, and for the most part, unpredictable. Anderson _et al_. (1976) remarked that the

size of the area clearcut and the proportion of the watershed clearcut have a great influence on the hydrologic response to forest treatment. Rich and Thompson (1974) found that removing mixed conifer forest vegetation in Arizona increased water yields approximately in proportion to the percent of area in cleared openings and in proportion to the amount of precipitation during the year. Hibbert (1967) plotted first year streamflow increases for forest treatments at Coweeta, North Carolina and Fernow, West Virginia, versus reduction in forest cover and showed that a linear relation between area cut and amount of yield increase existed. The simulation results for Jamieson Creek plotted in Figure 25 showed this linear relationship as well with the additional result that the relationship was directly proportional, i.e. doubling the area treated doubled the water yield increase. The results presented by Hibbert for Coweeta watersheds with northerly aspects demonstrate a similar, directly proportional relationship.

More difficult to compare is the magnitude of the overall increase in streamflow due to forest treatment because of the many factors which affect the result. Results from 10 watersheds, presented in Table 4, show that the average ratio of percent precipitation as annual streamflow on clearcut areas to percent precipitation as annual streamflow on uncut areas varied from 1.09 to 1.82. If only the Coweeta and Fernow watersheds are considered, the range for the same ratio is from 1.09 to 1.49 with an average of 1.22 for a variety of per cent area clearcut. If a similar analysis is done for the simulation results for

Table 4 Hydrologic response to clearcutting*

Watershed	Area (ha)	Location	Area Clear-cut %	% Precipitation as Streamflow		
				Uncut	Clear-cut	Ratio
Coweeta 13	16.1	N. Carolina USA	100	43	64	1.49
Coweeta 3	9.2	N. Carolina USA	100	33	40	1.21
Coweeta 22	34.4	N. Carolina USA	50	62	71	1.15
Fernow 1	29.9	W. Virginia USA	85	38	47	1.24
Fernow 2	15.4	W. Virginia USA	36	44	48	1.09
Fernow 7	24.2	W. Virginia USA	50	54	60	1.11
Wagon Wheel Gap	81.1	Colorado USA	100	29	36	1.24
Fool Creek	289.0	Colorado USA	40	37	48	1.30
Kamakia	35.2	East Africa	100	28	51	1.82
Kenya	688.0	East Africa	34	22	27	1.23
Mean	122.25		69.50	39.00	49.20	1.29

* Swanson and Hillman (1977)

Jamieson Creek the above ratio for 100% clearcutting for an evapotranspiration factor of 0.45 would be 1.08 for a daily evapotranspiration rate of 1.5 mm and 1.36 for a daily evapotranspiration rate of 5.0 mm. These results compare favorably with others reported in the literature. However, in view of the fact that at Seymour Falls six monthly precipitation totals are higher than the one used in the simulation and six are lower, one cannot safely say whether the simulated results would be representative of the annual effect at Jamieson Creek watershed or not. A long term per cent increase in flow due to clearcutting can only be obtained through the simulation of several years of data for Jamieson Creek.

9.1.5 Summary

The results just discussed show that the SLUICES model can successfully simulate observations of hydrologic response reported in the literature. The results for Jamieson Creek show that streamflow can be augmented through clearcutting and that the amount of increase in flow is directly proportional to the area clearcut. Of significant importance was the conclusion that the evapotranspiration level had a very significant effect on the increase in streamflow experienced, with a greater increase in streamflow being experienced from areas with a higher evapotranspiration level. Also the amount of evapotranspiration reduction on clearcut areas versus treed areas was shown to be a significant factor in determining the amount of increase in streamflow. Inclusion of various magnitudes of this factor would

allow for the simulation of the regrowth stages of clearcutting and the resulting reduction in streamflow. Simulation results showed that the greatest increases in discharge hydrographs were shown for the smaller sized storms.

Considering the simulations for clearcutting on Jamieson Creek using the SLUICES model, one can easily see why Hibbert (1967) would conclude that, based on a review of thirty-nine studies of timber harvest effects on streamflow throughout the world, the response to treatment was highly variable and basically unpredictable. These extreme variations in results can be rationalized considering the sensitivity of the results to the initial moisture content and the level of evapotranspiration, as shown by model simulations, as well as the probable wide variations in the physical characteristics of the watersheds studied.

9.2 Examination of the Contributing Area Concept Using the SLUICES Model

Because of its structure, its distributed nature, and its concentration on subsurface flow, the SLUICES model can have numerous hydrologic applications. One such application is the examination of the contributing area concept. This concept has essentially not been verified in the field because of the large extent of instrumentation required for such a verification. The SLUICES model, because of its distributed nature, routinely simulates the status of soil moisture in each element during the progress of a storm. Thus it should be ideally suited for examining the storm runoff process and some of the accepted

aspects of the contributing area concept.

9.2.1 Contributing Area

Traditionally infiltration approaches similar to that suggested by Horton implied that surface runoff is produced by rainfall excess which occurs at the ground surface when the rainfall intensity exceeds the soil's infiltration capacity. However, such techniques have not been found to be applicable on vegetated watersheds where overland flow is generally not experienced because the infiltration capacity exceeds the rainfall intensity. The absence of overland flow has suggested that storm runoff follows a subsurface route. However, this suggestion has created a dilemma to many hydrologists who could not rationalize the lack of overland flow and rapid hydrograph response on one hand and their belief in very slow subsurface flow, which obviously was occurring, on the other. Many current hydrologic models are still based on the assumption that the watershed is a lumped hydraulic system with streamflow being generated by processes which occur uniformly over the watershed surface. This suggests that the source area is equal to the watershed area.

Hewlett (1961) reported that the lower portions of a watershed normally exhibit higher moisture levels than upslope portions and would contribute to runoff earlier in a storm. Betson (1964) indicated that storm runoff usually originates from a small, relatively consistent part of the catchment. The Tennessee Valley Authority (1965) suggested that the watershed contributing area was a dynamic one and could vary in size during the course of a storm. Ragan (1967) showed that only a

small but variable portion of the watershed ever contributed flow to the storm hydrograph. Dunne and Black (1970) found that the importance of a hillslope as a producer of storm runoff depended largely on its ability to generate overland flow, also suggesting that storm runoff is generated on only a small portion of a watershed.

As a result of these research findings, a new runoff concept was suggested: that of contributing area. The concept hypothesized therein has been referred to as either variable source area, partial area, or dynamic watershed concept. This concept hypothesizes that only a relatively small portion of most watershed areas contributes direct runoff. The contributing area of a particular watershed is believed to be dynamic in nature and the properties of the contributing areas would vary from watershed to watershed according to topography, soil properties and vegetation. Furthermore, the contributing area would change with time within a watershed due to variable soil moisture conditions prior to and during storm events.

Dickinson and Whiteley (1970) introduced the concept of minimum contributing area, defined as the minimum area, which contributing 100% of the effective rainfall, would yield the measured direct runoff. They found the range of values to be extreme, ranging from 1 to 50%. However, the authors themselves wondered about the meaningfulness of the term in relation to the actual physical processes that occur on a watershed.

Riddle (1969) summarized values of contributing area found in the literature. These are presented in Table 5.

Table 5 Summary of Contributing Area Values Noted in the Literature*

Author	Catchment Area (km^2)	Catchment Characteristics	Contributing Area Characteristics	
Betson (1964)	0.015	Pasture cover +2% swamp	Mean value	4.6%
Betson (1964)	0.020	Area denuded of vegetation	Mean Value	85.8%
Tennessee Valley Authority (1965)	0.019	Heavily grazed pasture	Range	5 to 20%
Zovodchikov (1965)	1000 to 1500	Springmelt conditions	Range	20 to 60%
Ragan (1968)	0.460	Forested	Range	1.2 to 3.0%
Riddle (1969)	24	Agricultural; intermittent stream	Median value Range	2.2% 0.2 to 40%
Riddle (1969)	28	Agricultural; perennial stream	Median value Range	2.7% 0.5 to 8%

*Dickinson & Whitcley (1970), after Riddle (1969)

9.2.2 Formation of Contributing Areas

Return flow (water returning from beneath the soil to the ground surface) occurs whenever subsurface flow is unable to remove all the inflow. The consequent increase in the amount of water stored in the soil raises the level of saturation to the soil surface. Subsurface water can then emerge from the soil surface as return flow and proceed by an overland route. This concept is often used to help rationalize quick hydrograph response and apparent low subsurface velocities.

Overland flow causes the expansion of the perennial channel system. Subsurface flow, augmented by rain falling directly on the wetted area, feeds the expanding channel from below. The water table rises to the soil surface over an expanding area as rainfall progresses. This expansion of the channels takes place not only by the headward extension of the channels, but also by lateral expansion up the contiguous hillslopes. The saturated areas are generally located in valley bottoms and extend up slope during the wettest time of the year. The greatest probability of developing a saturated area exists on low-lying or concave areas with a considerable drainage area above it to supply seepage during a given storm. After the end of a storm, this saturated area contracts slowly as the soil drains and the channel shrinks back to its perennial length, at a rate dependent upon topographic and soil conditions.

9.2.3 Location of Contributing Area

Central to the contributing area concept is the belief that certain regions within a watershed contribute runoff to the storm hydrograph while other areas act as recharge or storage areas. Whether an area contributes to runoff depends on its physical position with respect to the channel, its soil properties and the storm characteristics. Generally valley bottoms are considered to be the areas that contribute to streamflow while ridge tops constitute recharge areas. The areas in between, often referred to as the dynamic zone, may be either contributing or recharging, depending upon the storm size and temporal characteristics, antecedent soil moisture contents and soil hydrologic properties. Figure 30a is a schmetic diagram of the concept as presented by TVA, while Figure 30b is a comparable one by Hewlett and Hibbert.

There are several reasons for wanting to know which areas of a basin yield saturation overland flow. The source of runoff is an important control of water quality. Knowledge of the areas that produce overland flow would permit the delineation of non-point sources areas of various contaminants. Better methods of predicting the location, magnitude and frequency of ground saturation at various times of the year would also improve land use plans which must take these areas into consideration. The watershed manager is faced with the need to identify the areas which have the greatest potential for hydrologic response to treatment. Thus, for flood prediction, water quality management

Figure 30a Dynamic watershed concept of runoff (T.V.A., 1965)

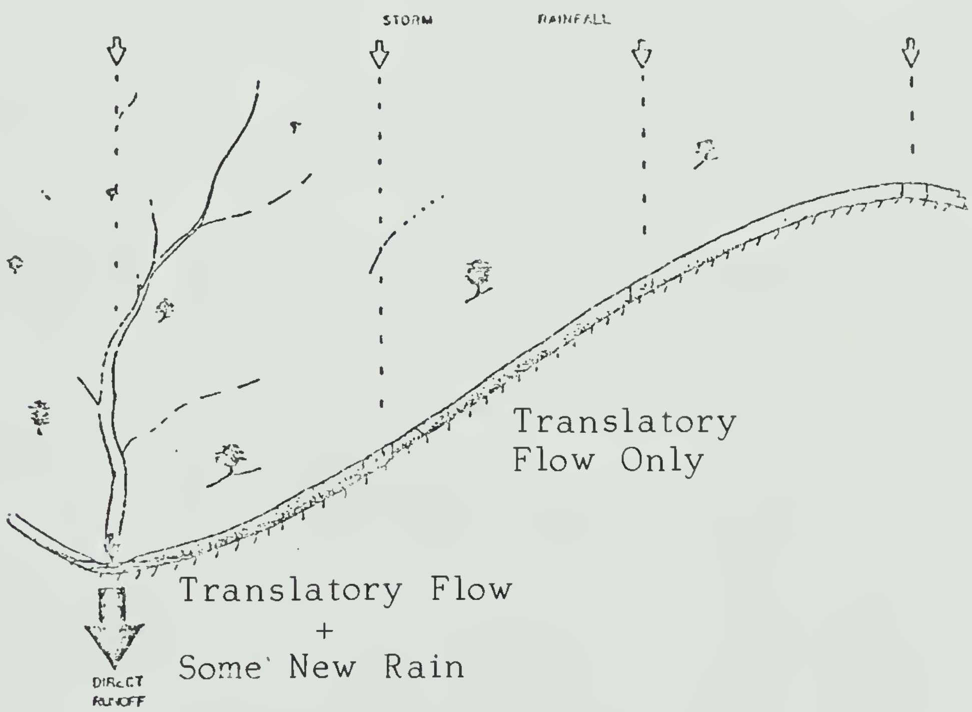


Figure 30b Channel expansion and source of subsurface stormflow (Hewlett and Hibbert, 1967)

and land use planning, routine methods for recognizing and quantifying the 'during storm' and seasonal variation of the saturated, runoff producing areas must be developed.

The nonhomogeneity of runoff response in a watershed can be attributed to variations in watershed soils, for example, variations in soil profile depth can cause a heterogeneous runoff pattern. As a result, projecting runoff data from one area to another is generally unsuccessful. No matter how similar two areas may appear, variations in the composition and depth of the various soil horizons can occur that may markedly influence how any particular area within the watershed will contribute to storm runoff. Clearly these soil variations must be considered before runoff data can be meaningfully projected.

9.2.4 Extent of Contributing Areas

Prediction of the extent of the variable source area is not an easy task however, because of the spatial and temporal variations displayed by it. Instrumentation requirements are extremely demanding for the monitoring of such areas on even the smallest of subbasins.

Soil morphology can be useful in indicating the distribution of saturated areas in a watershed. Soils subjected to seasonal waterlogging show grey-brown mottling or gleying. However, in saturated areas which exist for only a short period, gley morphology is unlikely to be very distinct in these zones. Variations in vegetative cover and land use practices may also complicate the effects on the water regime due to the morphology

of the surface horizons.

In some regions, it may be possible to use plants as general indicators of soil drainage and runoff producing zones. Unfortunately however, plants reflect only broad regimes of soil moisture and cannot give specific values at a particular season of the year.

At the present time there is no routine method for calculating the expansion of the saturated zone during a storm or of calculating the runoff produced by the processes described earlier. Detection of surface soil saturation seems to be the best criterion for identifying the source area of runoff. The SLUICES model automatically monitors the soil moisture status of each and every element of the watershed and thus is ideal for revealing, at any given time, which of the elements of the watershed have become saturated. Because of its distributed nature, the model is ideally suited for examination of the contributing area concept.

9.2.4.1 Extension and Contraction of the Channel Network During a Storm

Central to the contributing area concept is the expansion of the stream channel system to its maximum length at the time of the hydrograph peak and then a gradual recession in length. To investigate whether in fact model simulation showed that this was happening, the progress of the saturated area during the course of a storm given was followed. Storms 4, 8 and 11 were used for this purpose.

Figures 31a, 31b and 31c show which elements of the

watershed grid were saturated at various times during the storm. The results demonstrate quite clearly the expansion and recession of the saturated areas. Storms 8 and 11 resulted in similar peak discharges and display similar saturated area extent under simulation. Storm 4, of much lesser magnitude, displays only a modest extension of the saturated area. All trends are as one might expect but which one would have difficulty in quantifying.

The soil moisture status of each of the 300 elements in the Jamieson Creek watershed grid was examined at the time of peak flow for storms 11 and 4. The discharges were 8920 and 3875 m^3/h respectively. Figures 32a and 32b present the results for categorized soil moisture status for each of the elements. Categories for soil moisture status were set arbitrarily. As might be expected, the soil moisture status for the watershed was higher at the time of peak flow for storm 11 than for storm 4. At the time of peak discharge for storm 11, 42 elements or 14% of the watershed was saturated while at a similar time for storm 4, only 29 elements or 10% of the basin was saturated. Note as well that at the time of peak discharge, for both storms, every element showed some degree of soil profile saturation. Note that for storm 11, the greatest number of elements were in the 20 - 49% class while the greatest number of elements were in the 0 - 19% class for storm 4, again reflecting the wetter nature of the watershed at the time of peak discharge for storm 11 as compared to storm 4.

The contributing area concept also suggests a shrinkage of saturated areas during storm recession. Figure 33 shows the soil

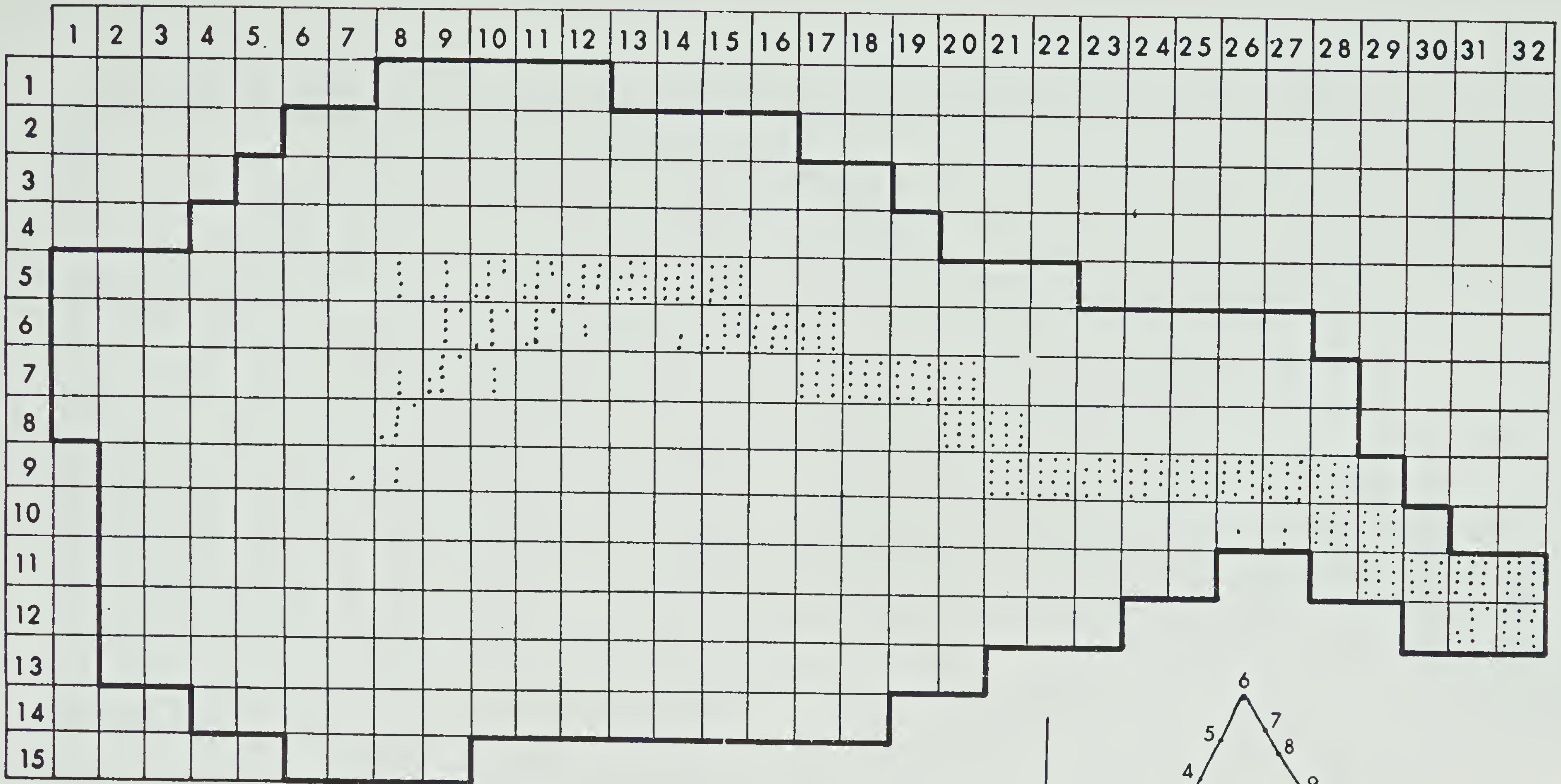
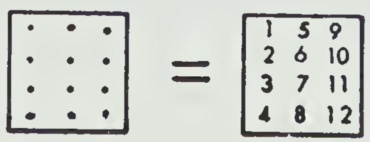
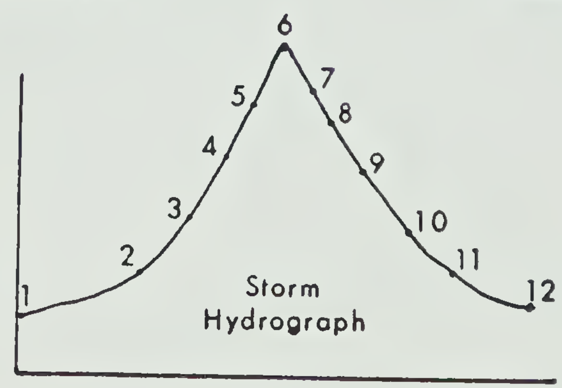


Figure 31a Saturated grid elements during the course of a storm
Storm 11





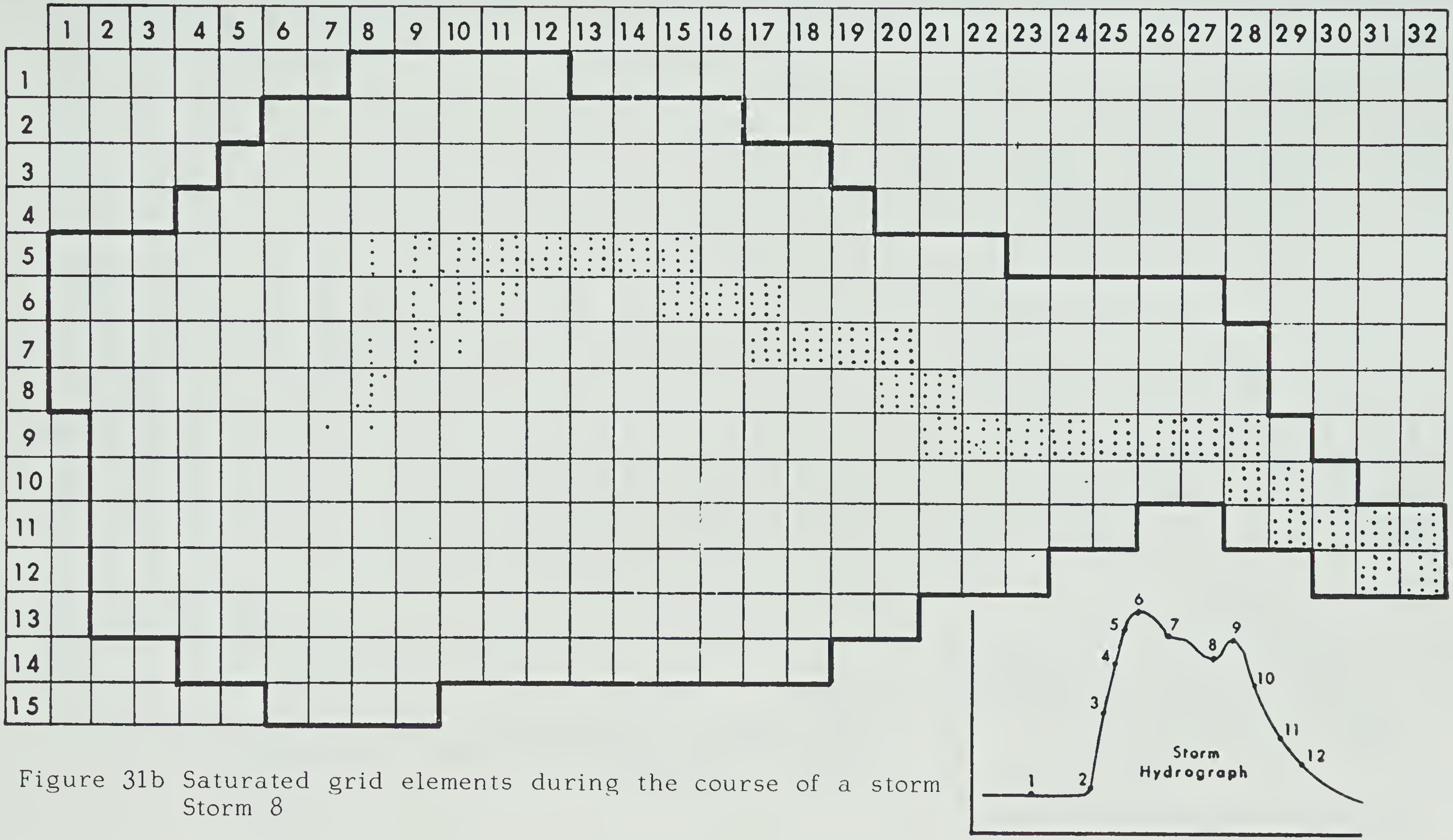


Figure 31b Saturated grid elements during the course of a storm
Storm 8

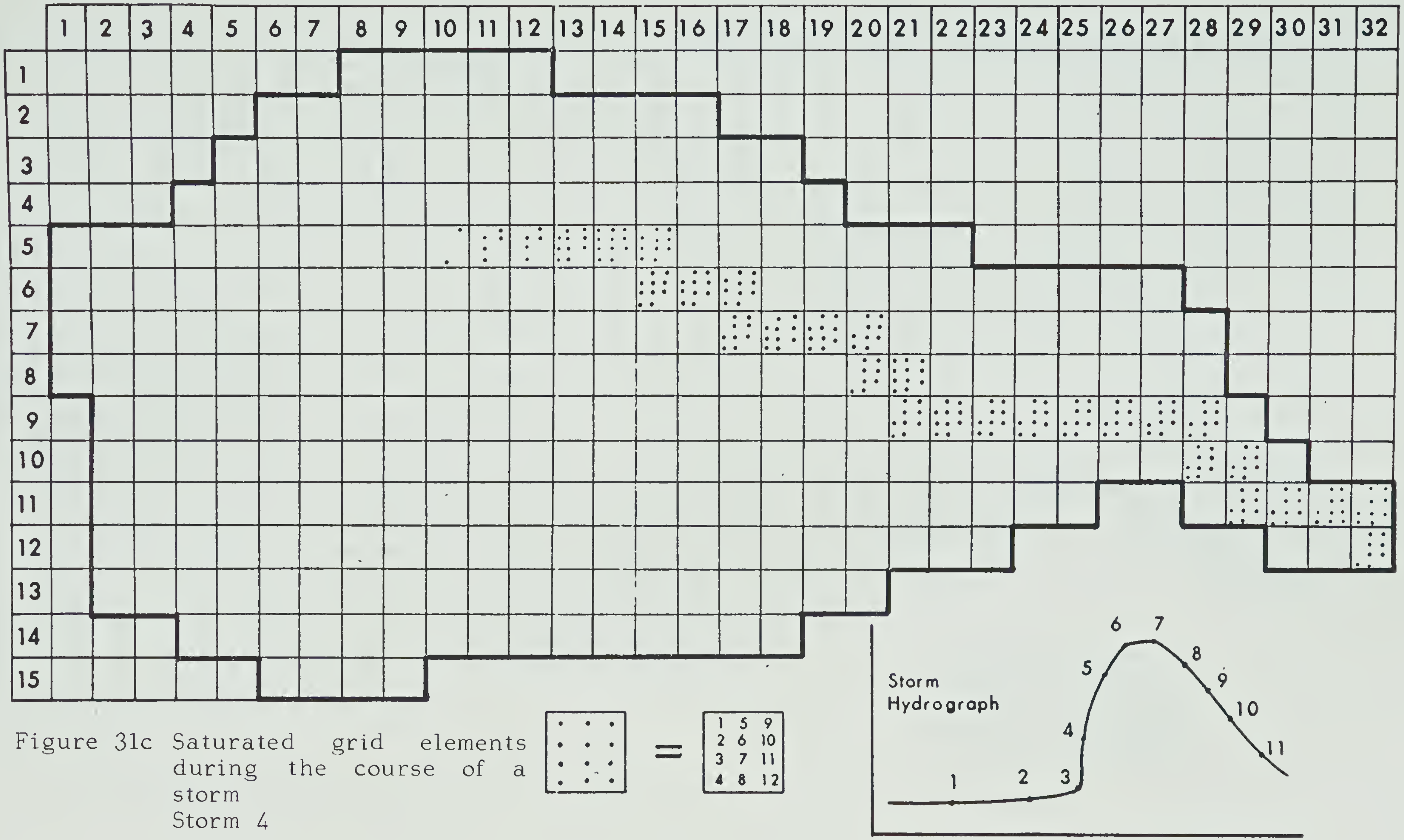


Figure 31c Saturated grid elements during the course of a storm
Storm 4

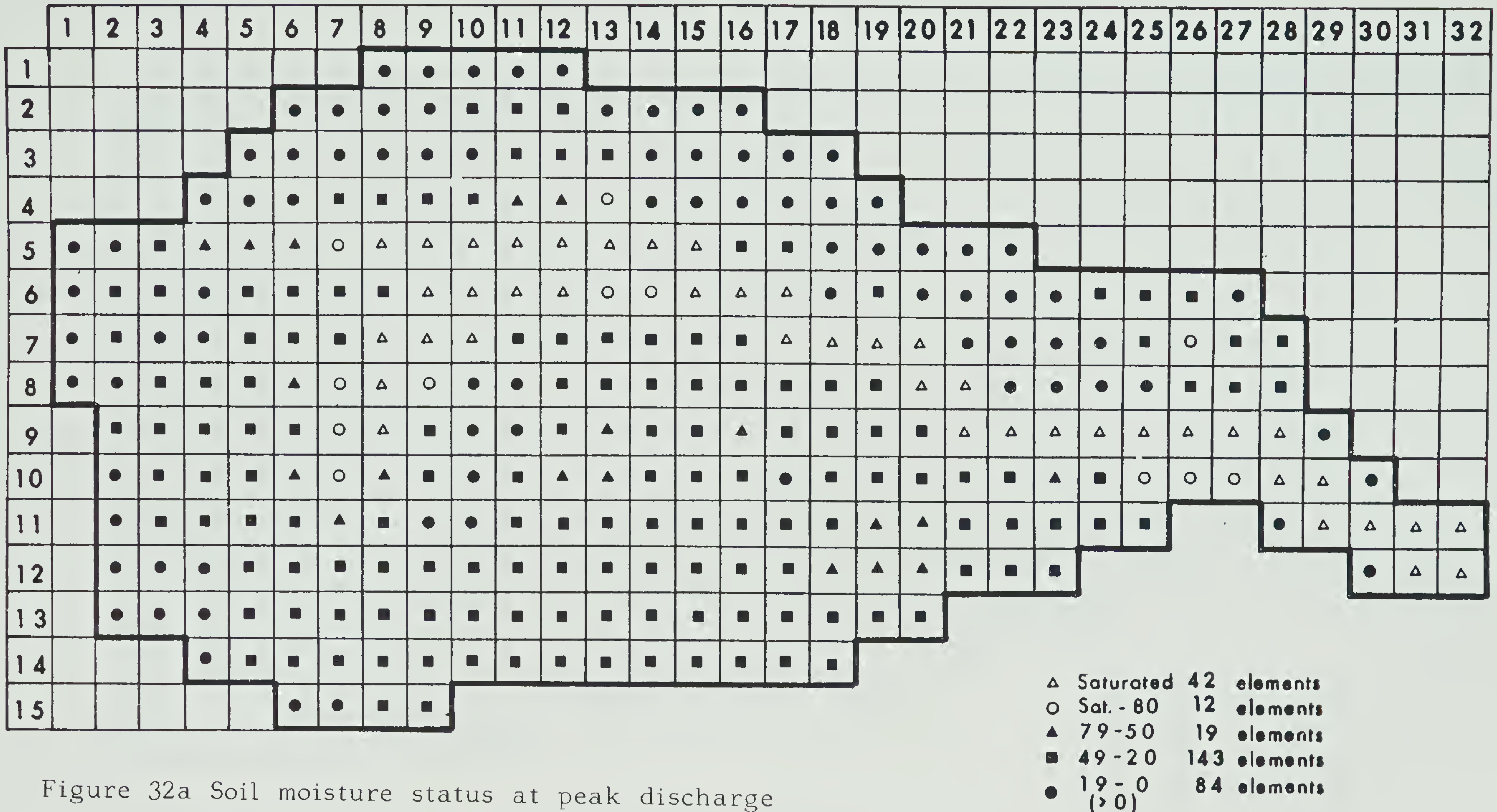


Figure 32a Soil moisture status at peak discharge
Storm 11 (8920 m^3/h)

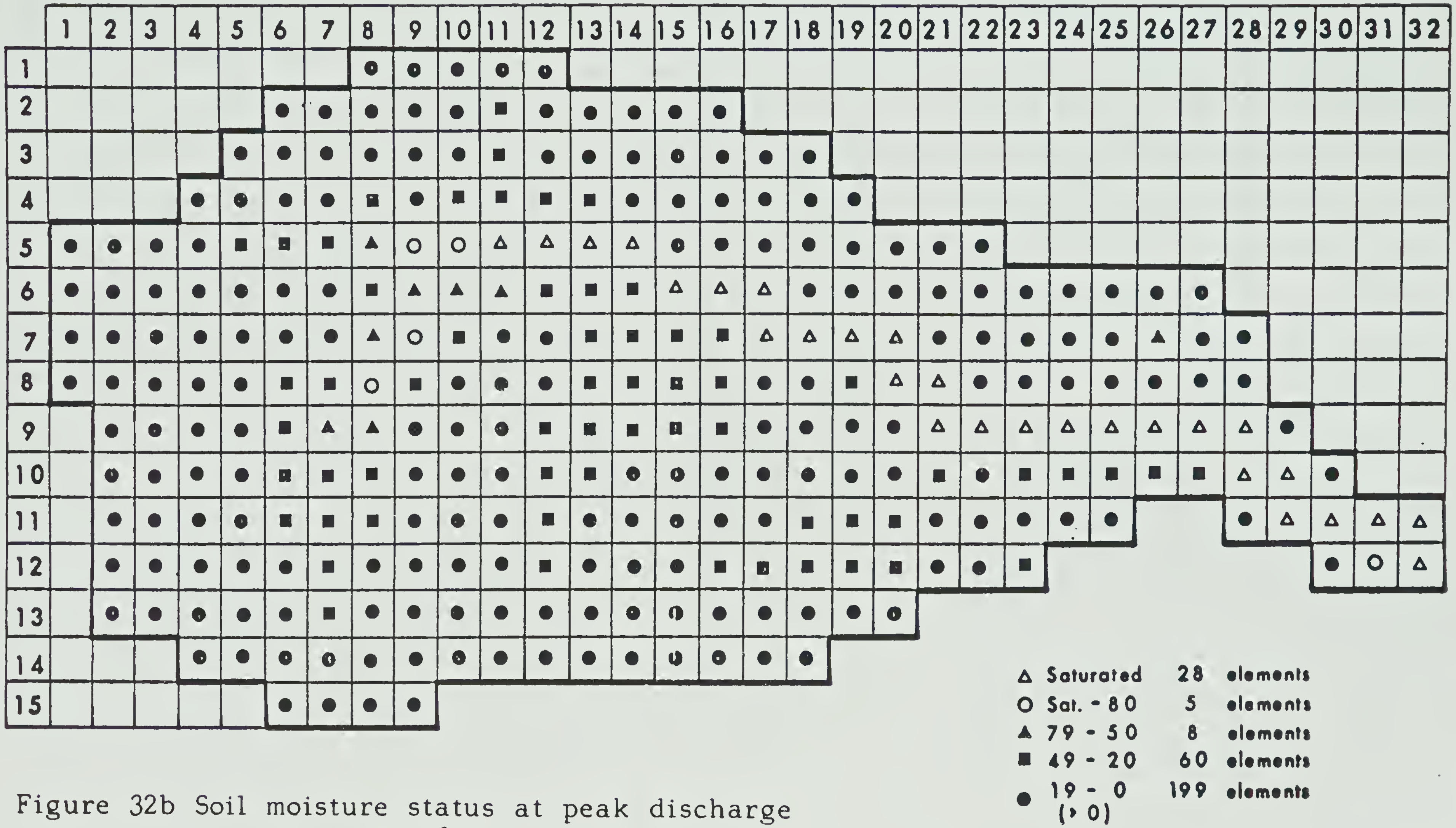


Figure 32b Soil moisture status at peak discharge
Storm 4 (3875 m^3/h)

moisture status of the watershed twenty hours after the peak for storm 11. (Compare with Figure 32a.) Twenty-seven elements or 9% of the watershed was saturated while only 21 elements showed any significant (1% or greater) soil profile saturation. Comparison of this figure with Figure 32a shows how the extent of the saturated areas in row 5 of the grid has shortened in length and how the 'channel' in column 8 is no longer present. Thus model simulations agree quite well with what the contributing area concept suggests happens during storm recession.

9.2.4.2 Upslope Expansion of Saturated Areas

To investigate the distributed nature of soil moisture along a slope during a storm, watershed column transects across the watershed were examined. Per cent soil profile saturation was then plotted for various slope positions on the transects for storms 4 and 11 at the time of peak discharge. The results are plotted in Figures 34a and 34b and certainly confirm the conceptual model of contributing area as presented by TVA (see Figure 14 for transect location). Generally speaking, at the time of peak discharge, per cent soil profile saturation decreases as one proceeds upslope from the low-lying areas. As well, per cent saturation was lower for all elements considered at the time of peak discharge for storm 4 as compared to storm 11.

9.2.4.3 Prediction of the Extent of Saturated Areas

In the introductory discussion on contributing area, the

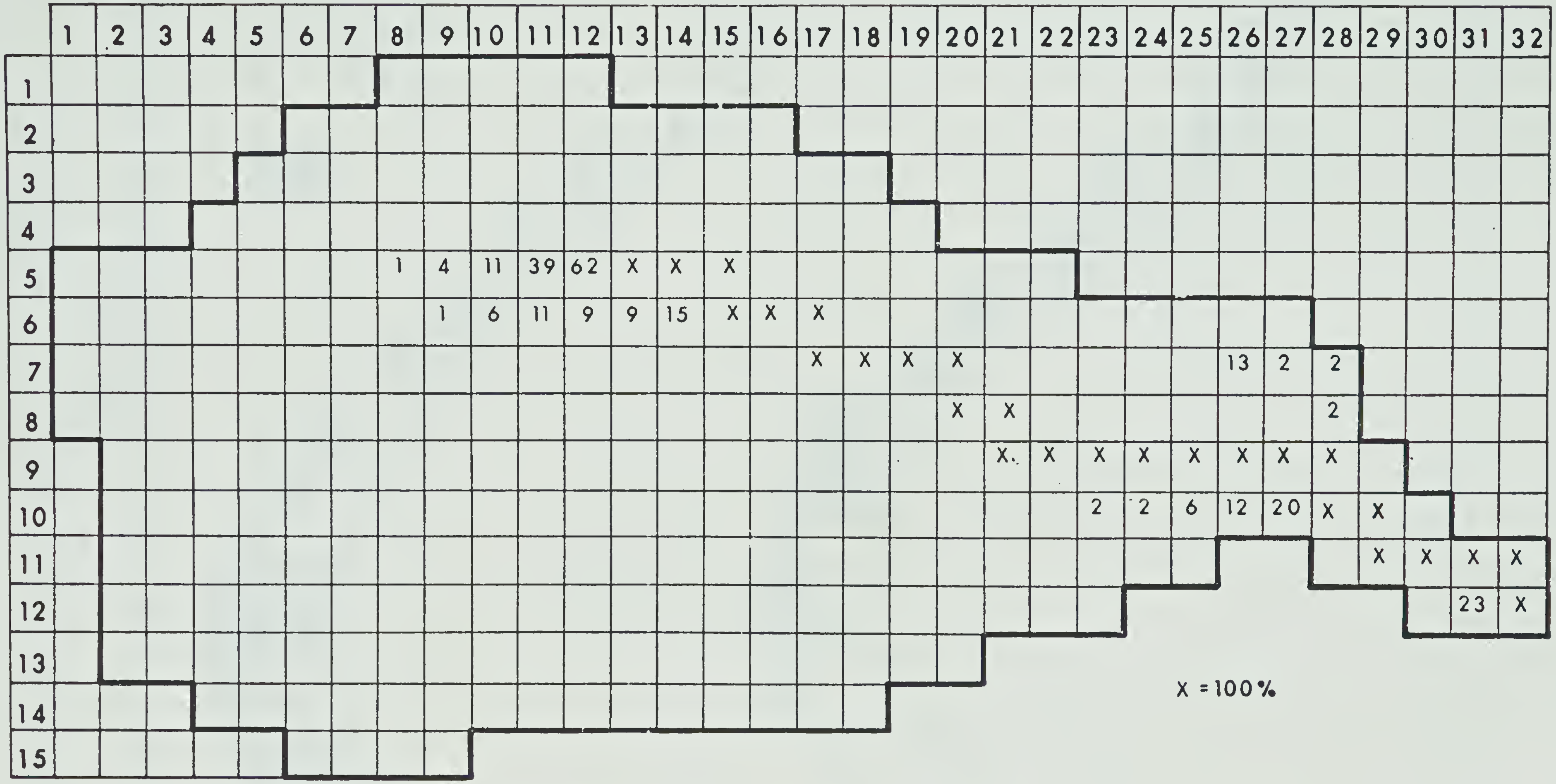


Figure 33 Per cent profile saturation during recession of Storm 11

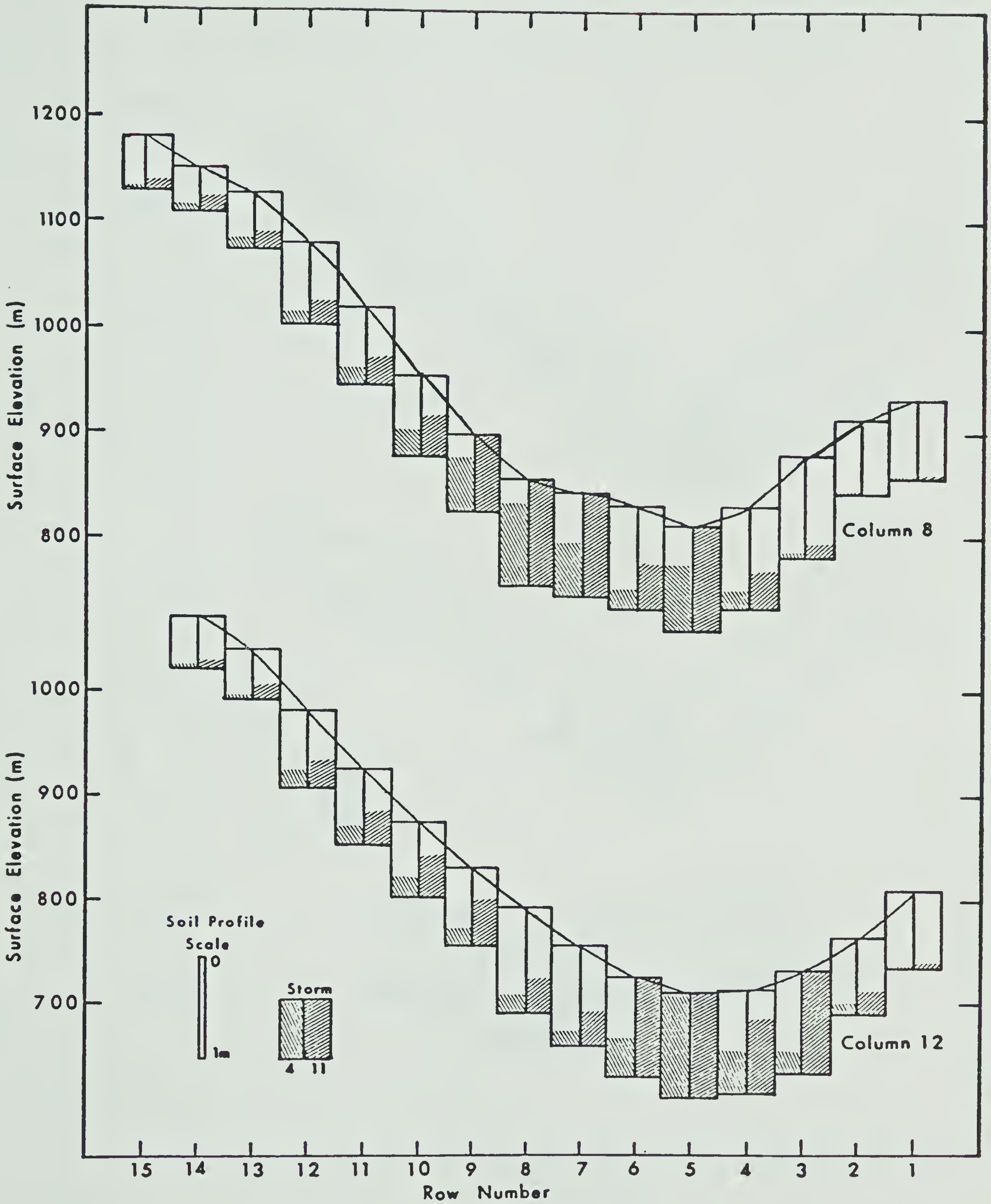


Figure 34a Profile saturation at time of peak discharge
Grid columns 8 and 12

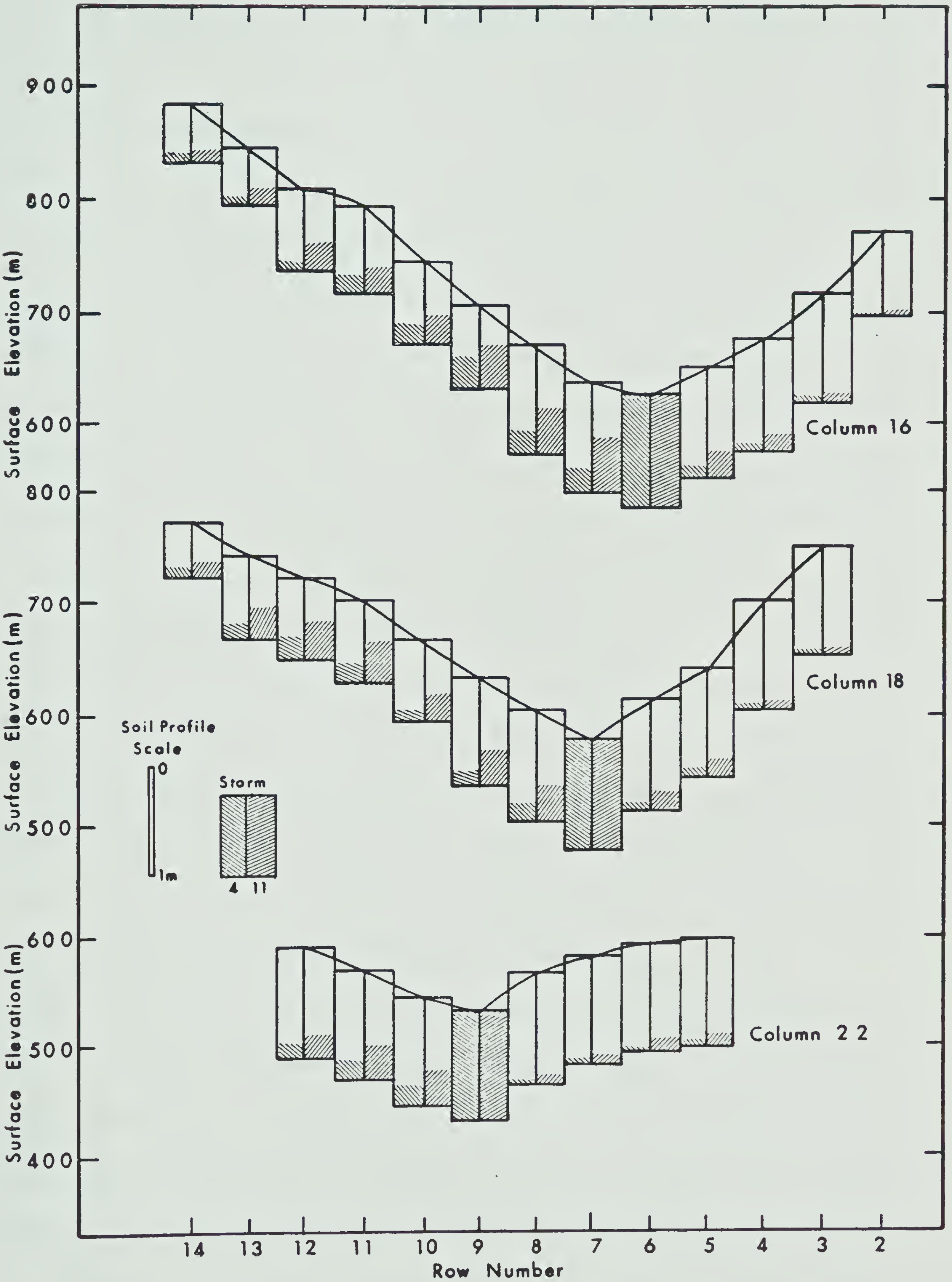


Figure 34b Profile saturation at time of peak discharge
Grid columns 16, 18 and 22

routine estimation of the saturated areas and their location was shown to be extremely desirable in the work of a variety of people. The capabilities of the model to predict routinely the location of the saturated areas has just been shown. In the estimation of the actual amount of saturated area, the soil moisture status for all elements was evaluated at various times during the discharge hydrographs for storms 4, 8 and 11. The number of saturated elements at given times were counted and the per cent saturated area was then compared against two easily obtainable hydrologic parameters: cumulative precipitation from the start of the storm till the time in question and the cumulative volume of discharge till the time in question. Figures 35 and 36 show the relationship between the simulated per cent saturated area and cumulative precipitation and per cent saturated area and cumulative volume of discharge respectively. The former relationship is excellent, and, the latter relationship is very good. These relationships show that, for this particular watershed, the per cent saturated area, as simulated, can be estimated from either cumulative precipitation or cumulative volume of discharge during the time of increasing discharge.

The values for saturated area range from approximately 7% to 14% of total watershed area. Unfortunately, as has been discussed previously, field verification of the results is an enormous task. Furthermore, comparison of the results from other watersheds, no matter how similar they appear to be, is of little value. However, the per cent saturated area values seem reasonable considering the physical nature of the watershed.

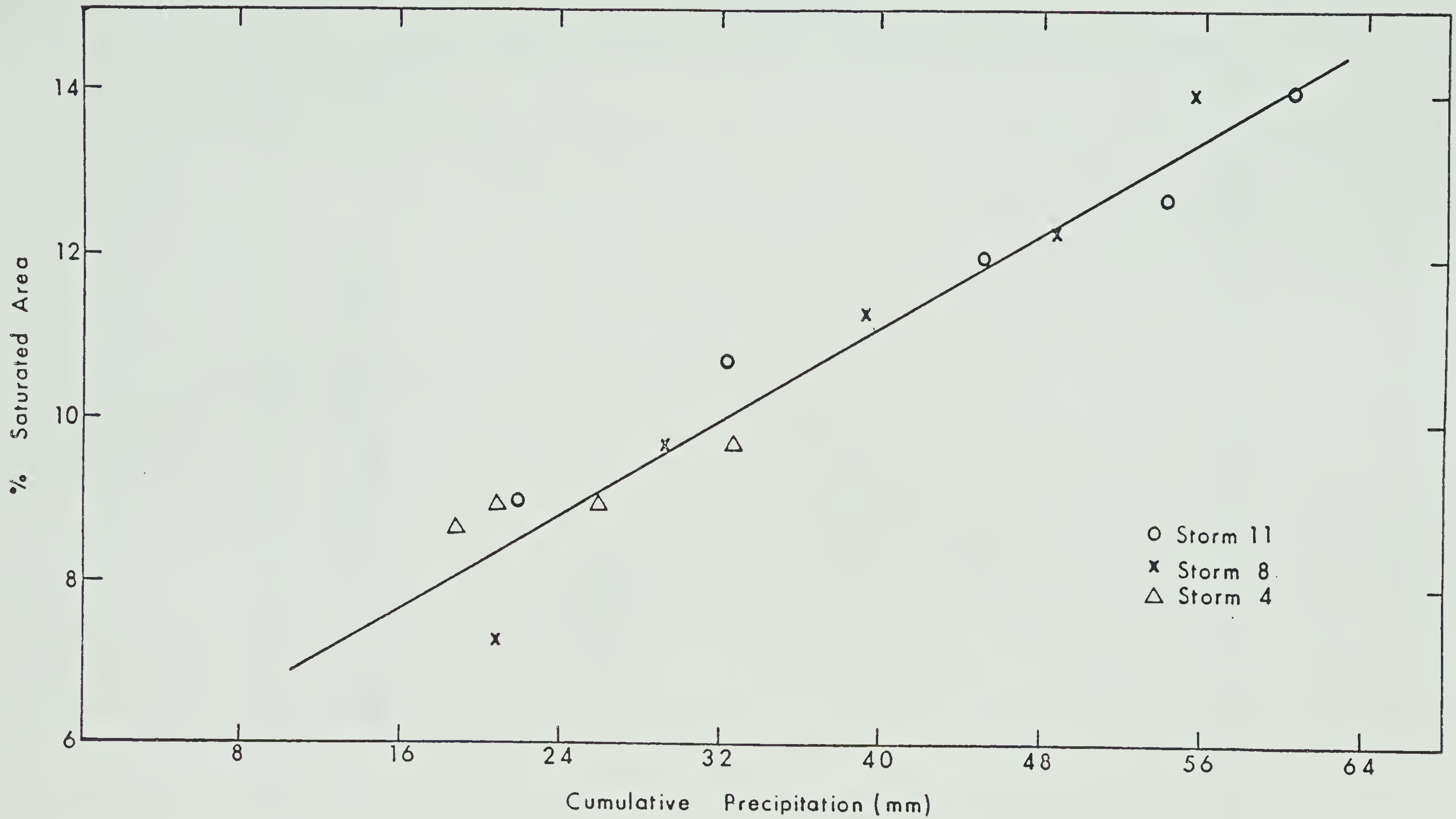


Figure 35 Per cent saturated area versus cumulative precipitation

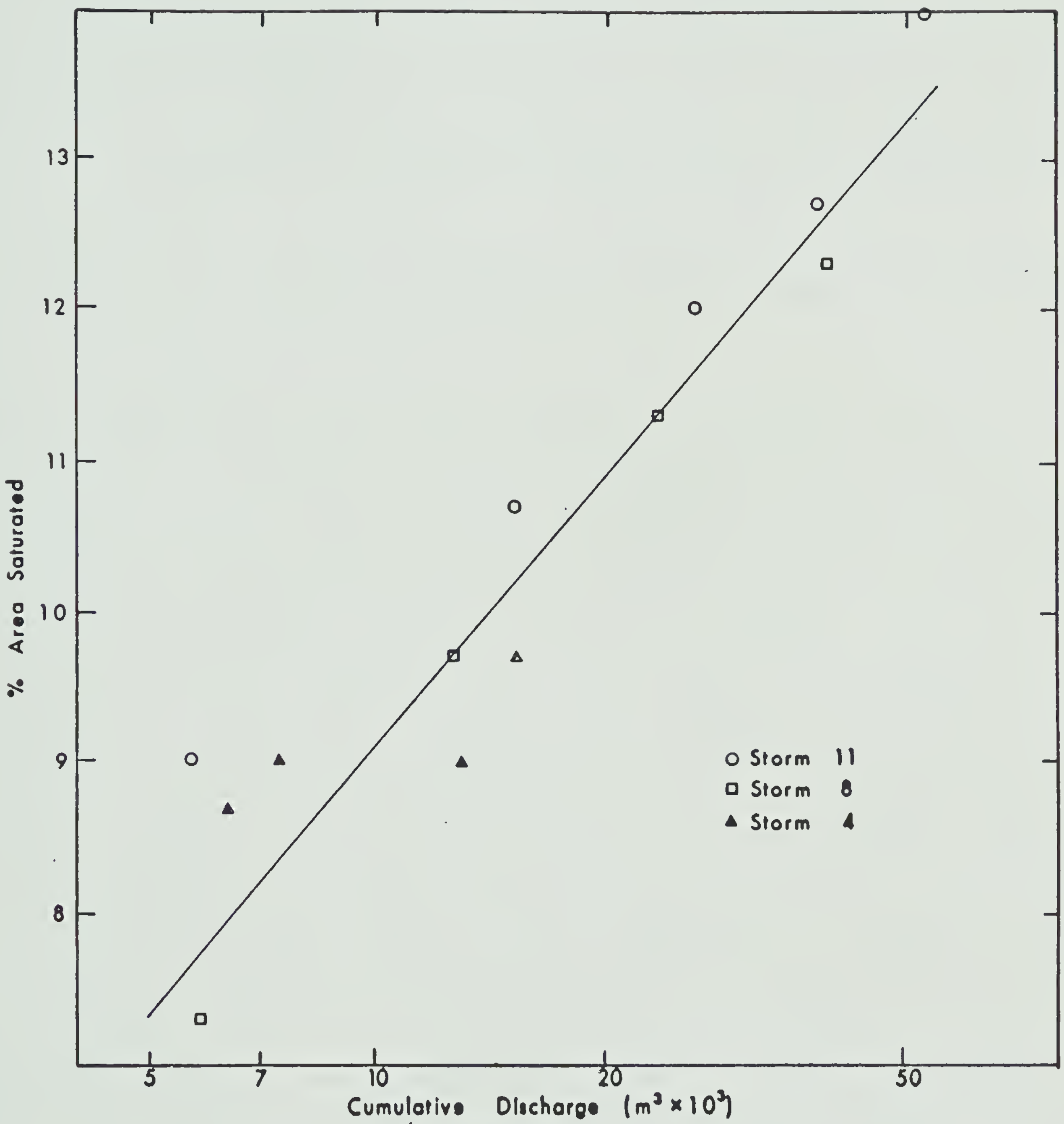


Figure 36 Per cent saturated area versus cumulative volume

9.2.5 Stream Channel Delineation

In the discussion of saturated overland flow, the formation of stream channels must also be considered. Leopold and Miller (1956) described three types of natural channels:

(1) the perennial stream: carries some flow all of the time,

(2) the intermittent stream: at low flow, dry reaches alternate with flowing ones along the stream length, and,

(3) the ephemeral stream: carries water only during storms.

This categorization takes into account the dynamic nature of the watershed as perceived by the contributing area concept. An hydrologic index commonly used to describe a particular stream network is drainage density, which is defined as the total length of stream channels per unit area of watershed. Its reciprocal (called the constant of channel maintenance by Schumm, 1956) gives an indication of the average distance between channels. Drainage density can be a useful index of the basin characteristics affecting the magnitude of streamflow from a drainage watershed. The relationship of actual discharge to drainage density, representing all types of flow, was shown to be of the form $Q \propto D_d^2$ (Gregory and Walling, 1968). The fact that drainage density varies with discharge in a particular watershed means that when drainage density and streamflow are related for different watersheds, consideration must be given to the level of streamflow and to the method used to derive the drainage density values. Because of the expanding and

contracting nature of the stream channel, drainage density within a particular watershed is not a static value. It is important to know to which discharge values the drainage density values, as derived from topographic maps, really relate. Map scale further introduces a problem as there are variations in the density of watercourses indicated on maps of various scales.

Two methods are presently being used for the determination of stream lengths and drainage density when data are obtained from topographic maps (Morisawa, 1957):

(1) Direct delineation of the streams on the map at the time of its making, or

(2) insertion of a stream into a drainage net whenever the V-shaped contours indicate a channel is present.

Thus, the usefulness of drainage density as a parameter is necessarily limited by the method used to delimit the drainage net and by the maps upon which the net is based. Langbein (1947) pointed out that the number of small headwater streams shown on maps (scale of 1:62,500) would vary with the season, wetness of the year in which the survey was made, as well as the judgment of the topographer and cartographer. Morisawa (1957) suggested that measuring stream lengths as printed on topographic maps is inaccurate and should not be used in quantitative work for watersheds of less than approximately 700 ha of drainage area. Schneider (1961) concluded that neither method of computing stream length, and subsequently drainage density, from topographic maps gives reliable estimates of the

true values as determined by field measurements.

9.2.5.1 Stream Channel Simulation

Since there should be a close agreement between saturated areas and stream location, results from the model's simulations were investigated to gain information about channel lengths and drainage densities. From the simulation runs for storms 4, 8, and 11, at a particular given discharge, the channel length was calculated by considering those elements which were saturated at that given discharge and summing their lengths. Results are given in Figure 37. The relationship is very similar to that reported by Gregory and Walling (1968), with the exception that their relationship did not show two separate sections. The upper segment of the relationship displayed in Figure 31 is the result of a channel branch forming in column 8 of the watershed grid beginning at a discharge of 4200 - 4500 m^3/h.

Drainage densities were calculated using the above calculated channel lengths and the watershed area. The relationship between discharge divided by area and drainage density is shown in Figure 38. The relationship again shows a similar shape to the one discussed above and is similar to that reported by Gregory and Walling (1968). The formation of the channel branch occurs at a drainage density of approximately 1.

The calculated drainage density values are quite low (varying from 0.87 to 1.23) as compared to the values (1.2 to 10.5) reported by Gregory and Walling (1968).

As an aid in assessing whether or not the simulated

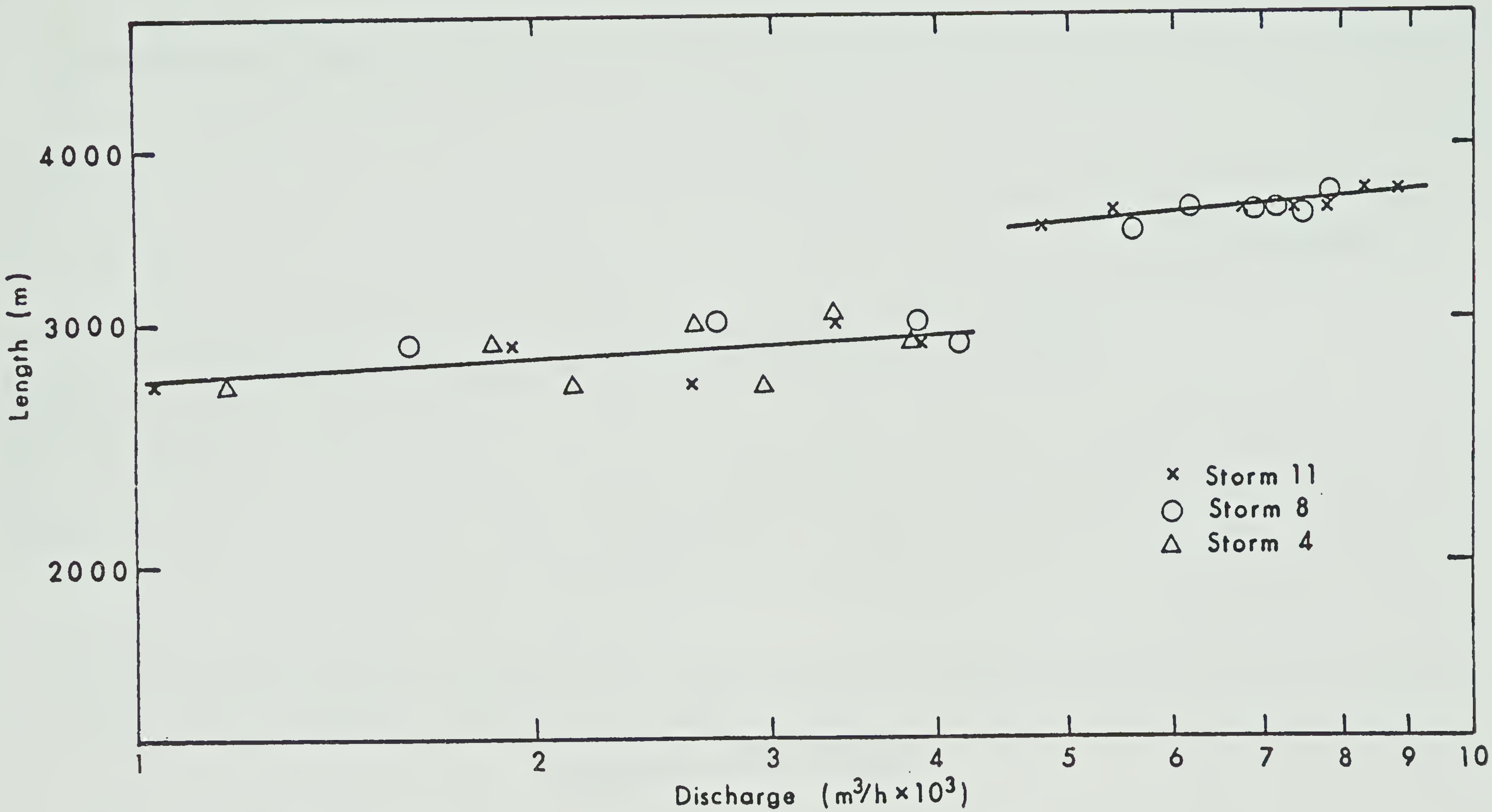


Figure 37 Stream length - discharge relationship

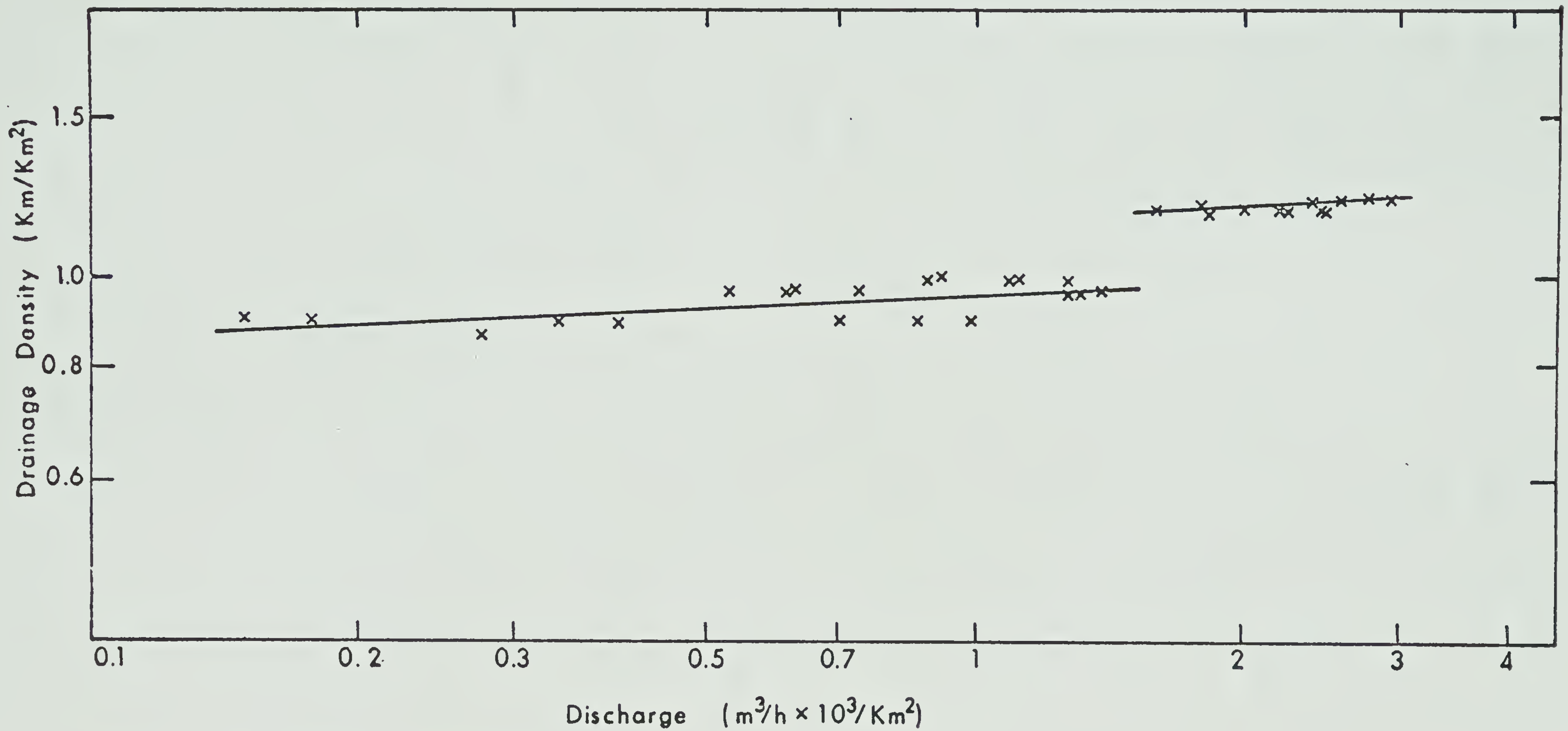


Figure 38 Drainage density - discharge relationship

channel lengths are meaningful, channel lengths were scaled from a topographic map of Jamieson Creek, scale 1" = 400'. The stream channel pattern is shown in Figure 39 and the stream lengths are listed for the various segments. The dotted lines correspond to intermittent channel sections. Note that stream channel sections (A - D) inclusive are 3020 m long, approximately the maximum length of simulated channel for discharges less than 4200 m^3/h. To check if simulated saturated areas correspond to near-stream channel areas, saturated elements as simulated at the time of peak discharge for storm 11, were overlain on the stream channel network as depicted on the topographic map (Figure 40). The agreement between simulated and actual channel areas is quite good. The close proximity of channel sections (C - D), (C - G), (G - H) and (G - I) has resulted in the entire area being saturated under simulation (see Figure 39). Channel section (B - L) did not appear in simulation. Examination of the grid elevations revealed that based on the input elevations, no channel could form there. Thus the absence of segment (B - L) is probably due to the inaccuracies of the input elevations (grid size too large) rather than model operation. Similar comments apply to the disagreement of part of the stream channel segment (A - B) and the simulated one.

Scaled Channel Lengths:

Segments	Length (m)
A-B	2060
B-C	205
C-D	755
D-E	1820
D-F	135
C-G	380
G-H	525
G-I	400
I-J	120
I-K	165
B-L	690

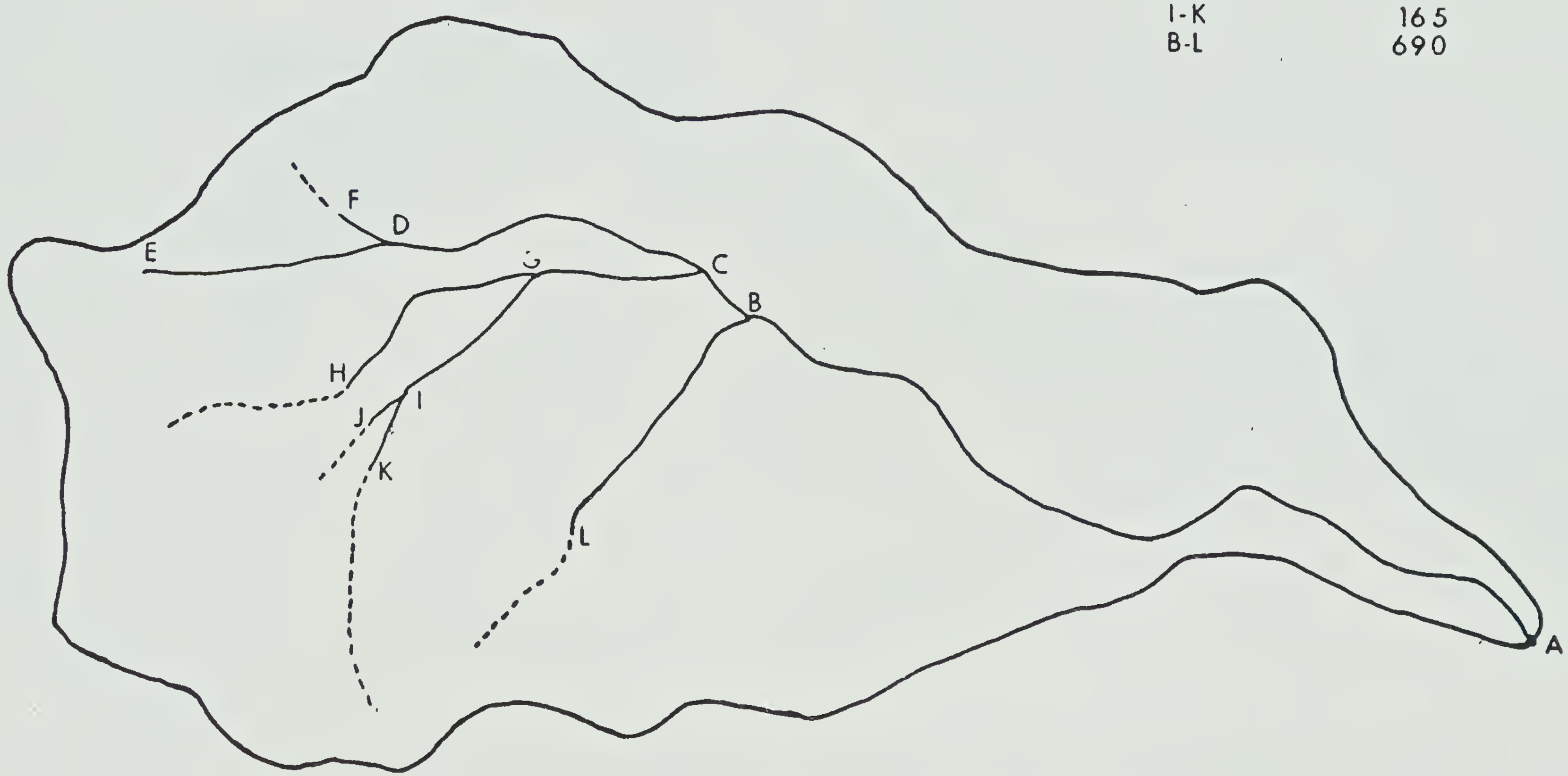


Figure 39 Stream channel delineation for Jamieson Creek

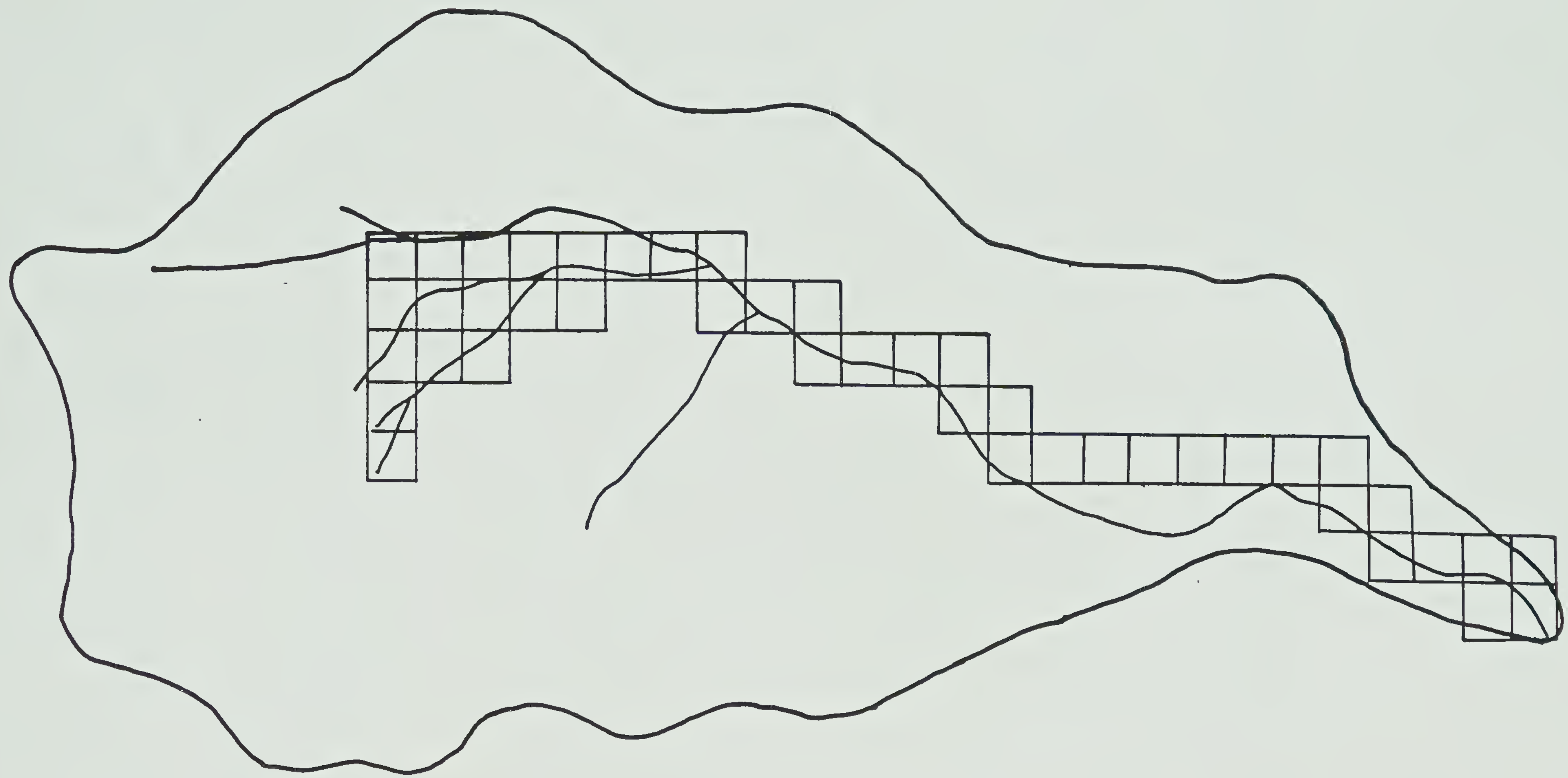


Figure 40 Comparison of simulated saturated areas and channel location

CHAPTER 10

SUMMATION

10.1 Summary

A distributed hydrologic model, the SLUICES model, was developed for the purpose of predicting the hydrologic response to land use manipulations, specifically timber harvesting. The model was developed for mountainous, forested watersheds where interflow plays an important role in stream discharge, where infiltration is not limiting and where a subsurface soil horizon impedes flow and, in conjunction with steep gradients, causes lateral flow. The soil profile was modelled in its classical, physical waterholding and transmitting sense and flow was calculated using Darcy's law. Overland flow occurred in instances where the soil profile became saturated from below. The distributed nature of the model was achieved through the use of a grid system having square elements, with the size of element being chosen by the modeller.

Model operation can be summarized as follows:

(1) Model development emphasized that simulation of soil moisture storage and withdrawal was realistic and representative.

(2) The model was able to represent watershed characteristics such as soil profile depth on a distributed basis. The model has the capability of

incorporating other soil properties, vegetative cover and precipitation on a distributed basis.

(3) Model structure was simple, containing only seven parameters. A sensitivity study revealed that the four most sensitive parameters were conductivity coefficient, saturation moisture content, field capacity, and drainage coefficient. Other parameters, such as wilting point and roughness coefficient, were less important.

(4) The model was applied to both a theoretical hillslope and an actual watershed.

(5) The model proved to be sensitive to antecedent soil moisture conditions and to evapotranspiration, both of which markedly influenced the hydrologic response to forest clearcutting of timber.

Qualitative conclusions drawn from simulated results for a hypothetical hillslope compared favorably with those reported by Freeze for his model. These conclusions pertained not only to process parameters but also to slope shape. One must note, however, that the SLUICES model is structured more simply than Freeze's model, but yet is capable of producing similar conclusions with much less data input required.

The SLUICES model was calibrated and validated for the Jamieson Creek watershed. Seven storms were used for simulation to assess the performance of the model on an individual storm basis. A 'dummy' storm was used to initialize antecedent soil moisture and streamflow conditions. Agreement between the

recorded and simulated discharge hydrographs for the seven storms tested was acceptable ranging from very good for storms with wet antecedent conditions to fair for storms with dry antecedent conditions. Thirty day simulations were then attempted with good results being obtained.

The model routinely simulates soil moisture for all elements. Results at the time of storm peaks showed that saturation of the soil profile, proceeding upstream, was indeed occurring. This result concurs with the contributing area concept. Also, because of its soil moisture simulation, the model can routinely show which elements had become saturated. Simulated results were compared with the mapped stream channel network of Jamieson Creek and again the agreement was most favorable. The expansion and contraction of the saturated areas was easily followed during the course of a storm and could be easily compared to other storms.

The predictive capabilities of the model with respect to watershed management for streamflow augmentation were assessed for different sized areas clearcut. The model showed that, based on the information available and the assumptions made, streamflow on Jamieson Creek could be increased by timber removal, with the amount of increase being proportional to the area clearcut. Because of its inherent structure, the model is also capable of providing predictions for the post-harvest response of a watershed as well. Accuracy of the predictions in this regard will be dependent on the availability of data for evapotranspiration levels during the regrowth period.

10.2 Conclusions

(1) The model itself was shown to be the ultimate limitation on the largeness of element size. Because of its nature, the model requires small elements for accurate results and the smaller the element size, the better the results. The smallness of element size, however, would be limited by the largest scale of topographic map available for the watershed. The influence of element size on the model operation should not prove to be limiting. Considering the basic nonhomogeneous nature of the watersheds for which the model was intended, large homogeneous areas would probably be the exception rather than the rule and therefore the choice of a smaller element size would be made anyway.

(2) Model parameters were found to be easily optimized within a range of physical possibilities. Only the optimized saturation value turned out to be identical to a measured value. The need for optimization of some model parameters, for example, field capacity and conductivity coefficient, is disappointing in view of the supposed physical nature of these parameters. The necessity for optimization for these soil parameters, however, can be due as much to the 'looseness' of the physical meaning of these parameters as to the idiosyncrasies of the model. The

optimized value for field capacity was, however, still well within physically acceptable limits.

(3) The most sensitive model parameter was found to be the conductivity coefficient. This finding parallels that of Freeze for his model. Conductivity coefficients equal to laboratory measured values of hydraulic conductivity were found to give uncharacteristically low hydrograph peaks during optimization for Jamieson Creek. Arguments have already been presented explaining the dangers of applying laboratory measured hydraulic conductivity measurements to field situations.

(4) The SLUICES model provided reasonably good simulation results for both specific storms and for continuous simulations of the discharge hydrograph for the Jamieson Creek watershed and proved capable of simulating the theoretical hydrologic response of forest clearcutting as reported extensively in the literature.

10.3 Suggested Model Refinements

(1) Because of the effect that model structure has on simulation results, further examination of the significance of the choice of grid size and time step is required before recommendations about such a choice can be made. Information of this nature would be useful in the application of the SLUICES model to other watersheds, particularly those for which larger

element size may be desirable.

(2) Further testing of the model for watersheds in different geographical locations will be necessary to evaluate the universality of application of the SLUICES model, which has been developed for rather specific conditions. Two groupings of test watersheds can be made in this respect. The first includes watersheds for which available physical data strongly suggest that interflow is the dominant flow, as in Jamieson Creek. Simulations on such watersheds would help establish the uniqueness or similarity of Jamieson Creek compared to other watersheds tested. The model should also be tested for watersheds where the physical characteristics of the watershed suggest that interflow should be the dominant flow but for which annual precipitation is much lower than for Jamieson Creek. These simulations would help establish the relative importance of interflow in such watersheds. The second grouping would include those watersheds for which interflow may not be the dominant flow process or those for which not enough physical data are available to help assess the relative importance of interflow. All of these applications of the SLUICES model would also provide information about the range of values of model parameters required for successful modelling of other watersheds.

(3) Because the model elements were developed with a planimetric view, it may be advantageous (and more realistic) to calculate the flow path lengths with the slope between elements being considered. This consideration would make model structure more complex and simulation more costly, but such calculations may improve simulation results.

(4) A more sophisticated technique of calculating actual evapotranspiration from potential evapotranspiration which includes plant factors as well as soil factors may be desirable if the model is to be applied to watersheds with a variety of cover type.

(5) A more detailed treatment of drainage water passing out of the soil profile would be required if the model was used in a study concerned with low streamflows. This treatment would probably involve formation of a deeper groundwater reservoir and corresponding hydraulic conductivity in the model to route the water that had drained out of the soil profile to the stream channel.

(6) An extension of the model to include multiple soil horizons rather than consideration of an average soil profile might be useful. This would allow for consideration of the litter layer in greater detail but would require characterization of parameters for each layer as well as drainage coefficients between layers.

(7) If simulations are to be made on an annual basis,

estimates of interception will have to be made and a technique of including it in the model will have to be devised.

10.4 General Research Knowledge Deficiencies as Shown by Model Use

By its very nature, development of any hydrologic model will call upon the knowledge of a variety of scientists and engineers. Development of the SLUICES model demonstrated several deficiencies in general hydrologic knowledge whose removal would make modelling easier and more meaningful. Some of these deficiencies are:

(1) An acceptance that the outright application of laboratory measurements of hydraulic conductivity to field applications, particularly in nonhomogeneous soils such as forest soils, can be extremely misleading and that these values should be used only as first estimates. Mosley (1979), using tracer studies, has shown that actual flow rates can be at least two orders of magnitude higher, due to the inclusion of root channels and animal burrows. The resulting flow pattern does not fit into the pattern associated with the classical definition of hydraulic conductivity. More research is required on this type of flow as is an evaluation of how limiting a Darcian type approach to this type of flow calculation can be. Present equations for turbulent flow are too rigorous for field application. Emphasis should be placed on the

development of a simple formula which might account for non-Darcian flow but yet would be based on easily measured parameters in the field.

(2) Data on the magnitude of evapotranspiration reduction in partial clearcut areas within completely forested areas are crucial to proper modelling of hydrologic response; yet, in many studies evapotranspiration is calculated as a residual term in a water budget or is lumped in a 'loss' term. Direct measurements of evapotranspiration are costly to obtain in the field but reliable estimates of water yield increases cannot be made until such data are available.

10.5 Research Areas for Further Model Application

Because of its distributed nature and its emphasis on subsurface flow, the SLUICES model may be useful in locating watershed areas where slope stability may become a problem if vegetation in those areas is removed.

Similarly the model may be useful as a tool for assessing forest site indices and delineating areas which would become saturated as a result of vegetation removal causing subsequent vegetation growth problems.

The full extent of the capabilities of the SLUICES model, offered by its distributed nature, were not utilized due to a lack of detailed extensive surficial geology and soils information for the watershed used for model validation. However, even in instances where relatively little data is available, the model can

be instructive in comparing theoretical results for proposed management practices on a relative basis. The model could also, in these instances, be used to guide field research work.

Watersheds experiencing markedly distributed precipitation patterns could be easily modelled if suitable data on these patterns were available.

The SLUICES model could be applied to watersheds which have a significant portion of their precipitation as snowfall if a snowmelt subroutine were added to the model. A snowmelt subroutine similar to that utilized by Leaf and Brink (1973) could be incorporated. The distributed nature of the model would prove most appropriate for those watersheds which have only small clearcut blocks to help in snow entrapment.

The SLUICES model could also be applied to agricultural watersheds with the addition of an infiltration function, as equations similar to the type devised by Holtan and based on soil water storage could be quite easily incorporated into the model. Soil moisture redistribution could be quantified through the use of a conductivity coefficient which is a function of soil moisture. The distributed nature of the model would allow it to handle areas of different infiltration capacities due to soil or vegetation differences. Non-uniform cover types, usually found in agricultural watersheds, could be handled as well. For example, combinations of cultivated areas, grassland and forest cover could all be included.

The concept of contributing area is one often discussed in the literature but for which little quantitative information is

available. Application of the model to an area where soil saturation could be monitored would help assess its apparent ability to predict the expansion of these areas.

Application of the model to watersheds that have been clearcut, such as those of Coweeta or Fernow would be desirable to assess quantitatively the capabilities of the model for predicting water yield increase due to clearcutting. Changes in infiltration capacity and/or water holding capacity of the soil due to harvesting techniques may now have to be considered to achieve better simulations. These changes would probably occur on a distributed basis as well, but could be easily handled by the SLUICES model.

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APPENDIX

APPENDIX 1

SOIL PROFILE CHARACTERIZATION

Soil is defined as the naturally occurring unconsolidated mineral or organic matter, at least 10 cm thick, that occurs at the earth's surface and is capable of supporting plant growth. Soil is a collection of naturally occurring three-dimensional bodies with vague limits that differs from geological materials as a result of soil forming processes (Canada Soil Survey). When an exposed cross-section, as in road cuts or excavations, is viewed, one or more soil layers are usually evident. These layers often parallel the earth's surface and are called horizons. Horizons differ from each other in properties such as color, structure, texture and chemical, biological and mineralogical composition. The vertical sequence of horizons which is observed for any soil is termed the soil profile.

The major mineral horizons are designated as A, B, and C and the major organic horizons are L, F, and H (mainly forest litter at various stages of decomposition), and O (derived mainly from bog and swamp vegetation). Subdivisions of horizons are labelled by adding lower case suffixes to some of the major horizon symbols, for example, Ah or Ae, and further characterize the respective horizon. General descriptions of the major horizons follow:

(1) Surface organic layers: In forest soils these layers comprise the surface accumulations of foliage and

other plant debris. They can often be separated into zones, vertically, according to the degree of decomposition. Three layers are usually specified:

L : slightly decomposed material, readily identifiable as to origin,

F : partially decomposed material, origin difficult to ascertain,

H : origin unidentifiable and the lower portion mixed in with the mineral soil.

(2) A Horizon: This is a mineral layer found at or near the surface, in the zone of leaching or eluviation of materials in solution or suspension or of maximum in situ accumulation of organic matter or both. The accumulation of organic matter is usually expressed morphologically by the darkening of the soil surface (Ah) and conversely the removal of organic matter is expressed by a lightening of the soil color usually in the upper part of the profile (Ae). The removal of clay from the upper horizon results in a coarser soil texture relative to the underlying subsoil layers.

(3) B Horizon: A mineral horizon below (1) and (2) characterized by the accumulation of one or more of silicate clay, iron, aluminum, or humus. Clay accumulation is indicated by finer soil texture and by clay coating peds and lining pores (Bt). Soil structure with coatings or stainings and significant amounts of exchangeable sodium is designated Bn with other changes in structure from that of the parent material Bm.

(4) C Horizon: A mineral horizon exhibiting an accumulation of carbonates and soluble salts and characterized by cementation. This horizon usually overlies consolidated bedrock.

L F H — Organic horizon, which may be subdivided into: L (raw organic matter), F (partially decomposed organic matter), and H (decomposed organic matter).

A — A mineral horizon at or near the surface. It may be a dark-colored horizon in which there is an accumulation of humus (Ah), or a light-colored horizon from which clay, iron, and humus have been leached (Ae).

AB — Transition horizon.

B — Mineral horizons that (1) may have an enrichment of clay (Bt), iron (Bf), or organic matter (Bh); or (2) may be characterized by a columnar structure and a significant amount of exchangeable sodium (Bn); or (3) may be altered to give a change in color or structure (Bm). Usually lime and salts have been leached out of this horizon. (4) The symbol (j) is used with the above suffixes to denote a failure to meet the specified limits of the suffix.

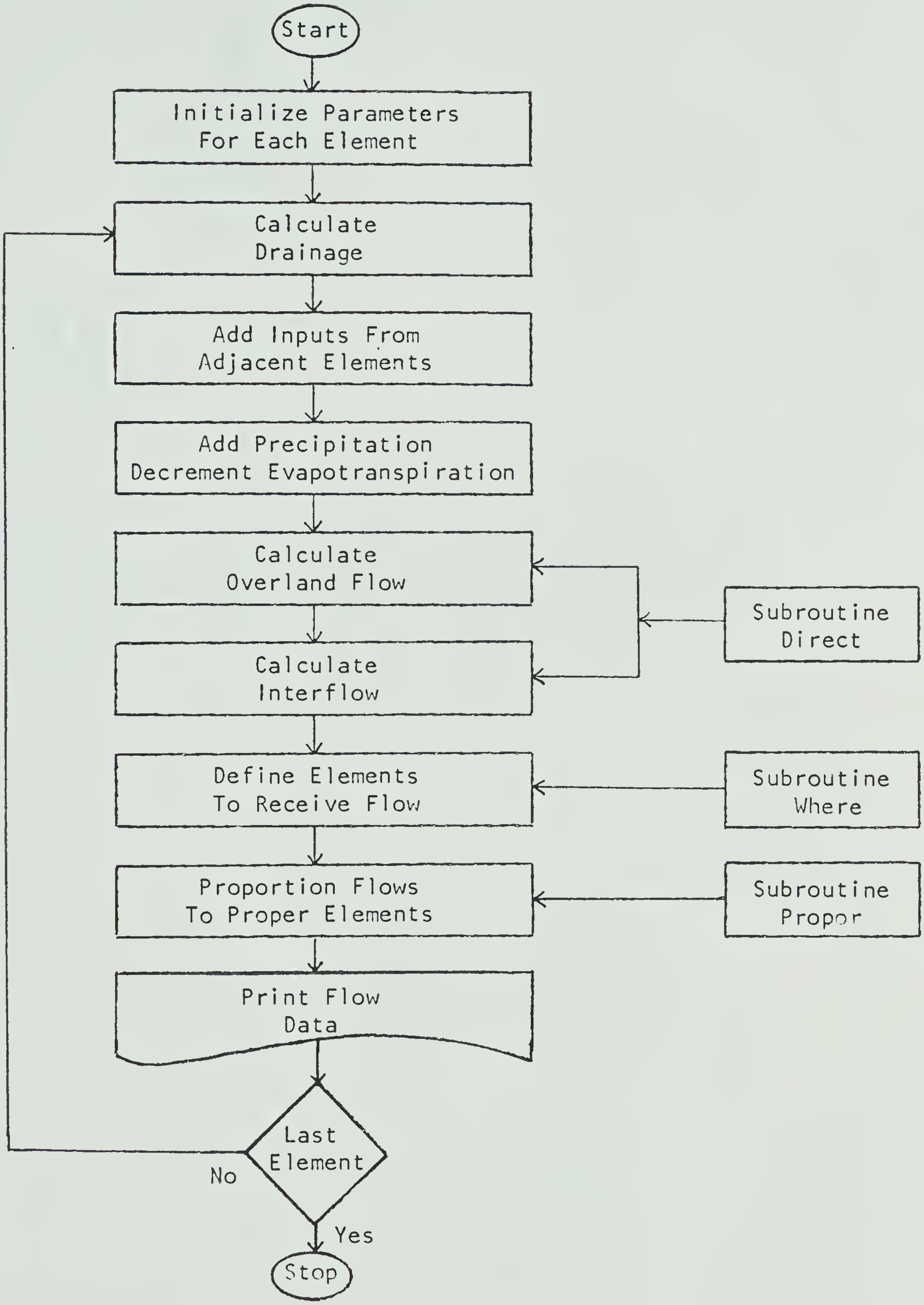
BC — Transition horizon.

C — Mineral horizon comparatively unaffected by the soil-forming process operative in the A and B horizons except for the process of gleying (Cg) and the accumulation of calcium and/or magnesium carbonates (Ck) and soluble salts (Cs).

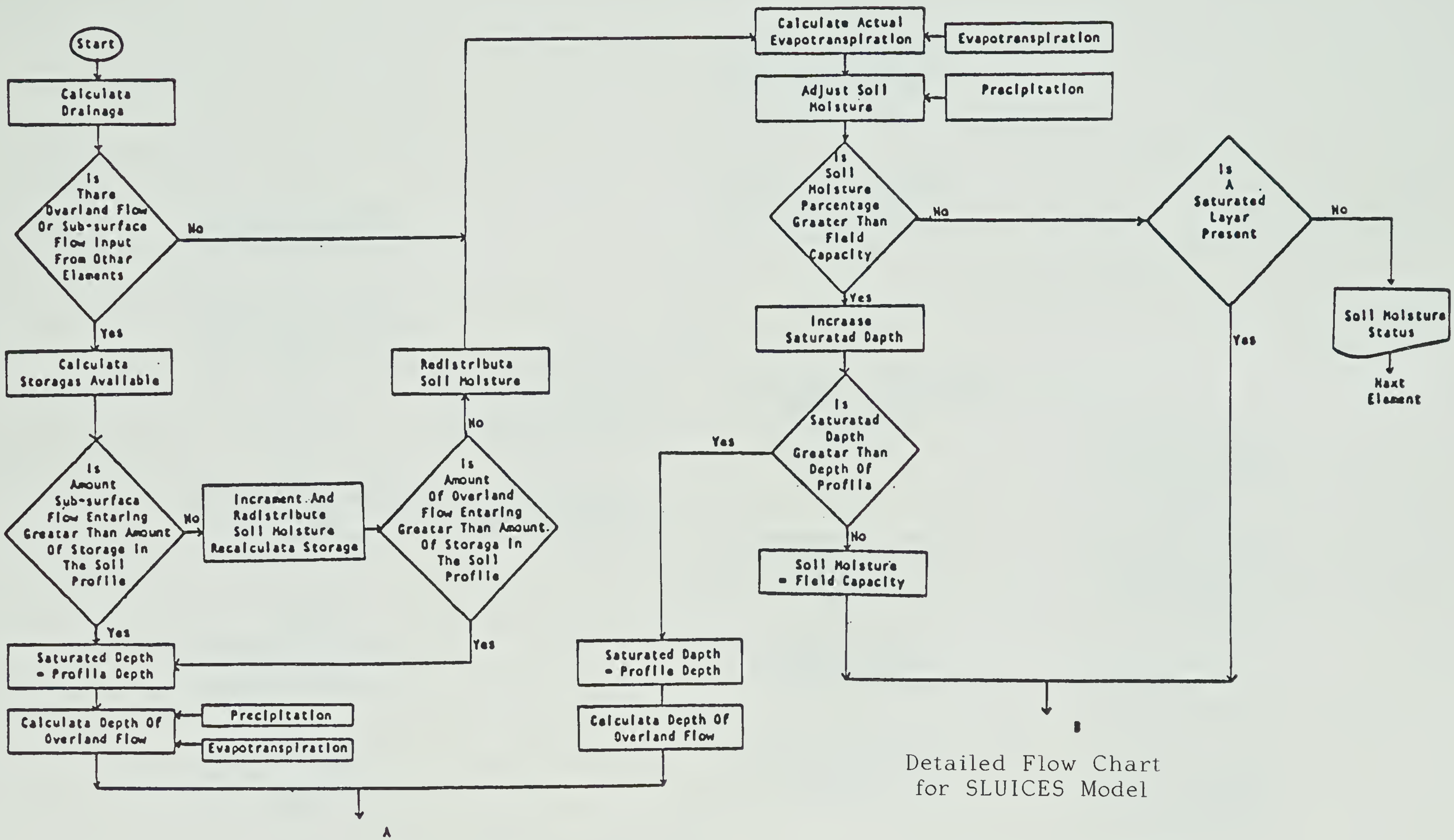
Schematic diagram of a soil profile showing various horizons.

APPENDIX II

MODEL FLOWCHARTS

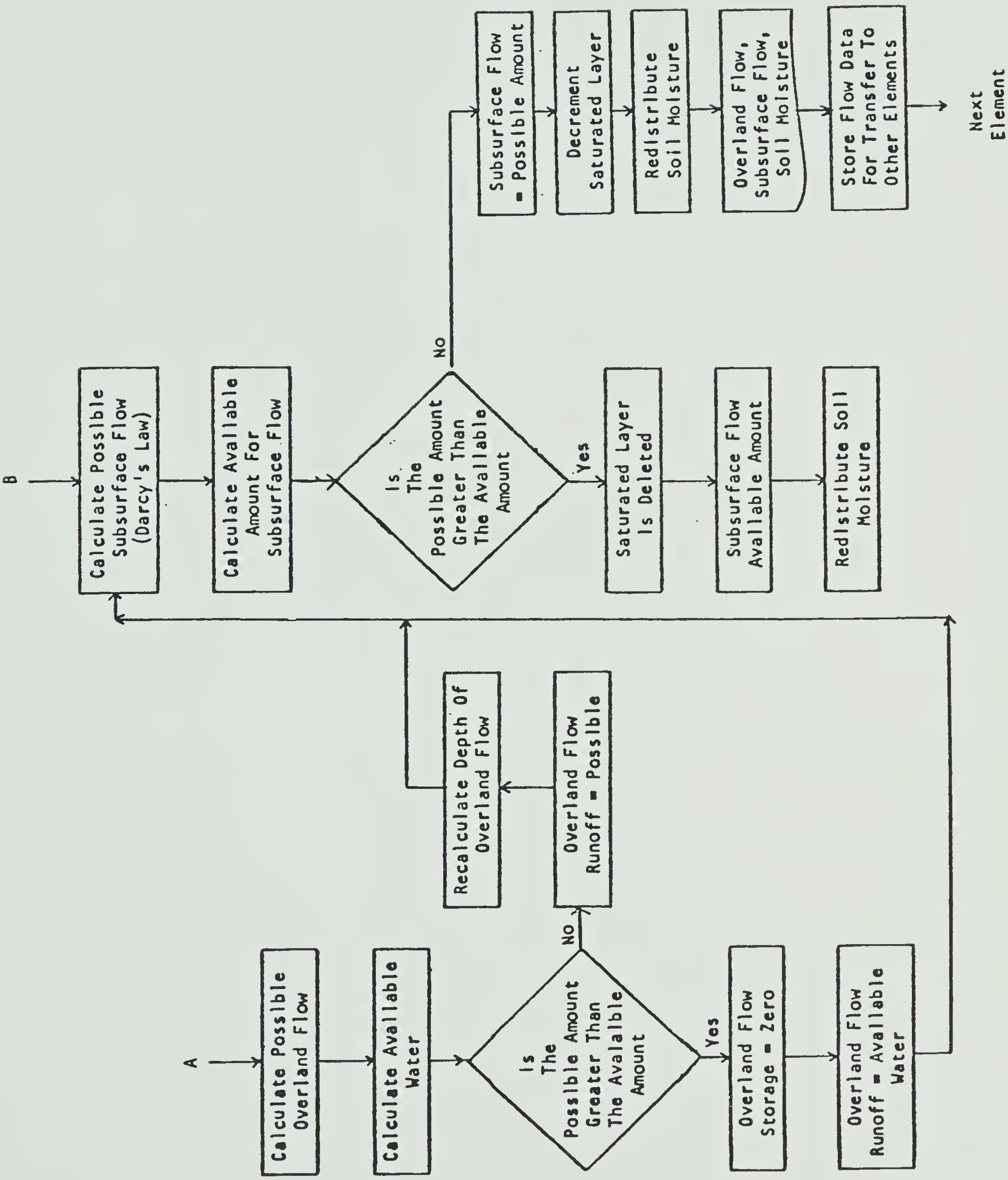


General Flow Chart for SLUICES Model



Detailed Flow Chart
for SLUICES Model

(continued on next page)


A
Calculate Possible Overland Flow
Calculate Available Water
Is The Possible Amount Greater Than The Available Amount
No
Overland Flow Runoff = Possible
Recalculate Depth Of Overland Flow
Yes
Overland Flow Storage = Zero
Overland Flow Runoff = Available Water
B
Calculate Possible Subsurface Flow (Darcy's Law)
Calculate Available Amount For Subsurface Flow
Is The Possible Amount Greater Than The Available Amount
Yes
Saturated Layer Is Deleted
Subsurface Flow Available Amount
Redistribute Soil Moisture
No
Subsurface Flow = Possible Amount
Decrement Saturated Layer
Redistribute Soil Moisture
Overland Flow, Subsurface Flow, Soil Moisture
Store Flow Data For Transfer To Other Elements
Next Element